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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 64, PP. 381-390, 2 FIGS. APRIL 1953

TERMINOLOGY FOR STRATIFICATION AND CROSS-STRATIFICATION IN SEDIMENTARY ROCKS

By Edwin D. McKee and Gordon W. Weir

Abstract

A terminology is suggested to aid the field geologist in describing the structures of stratified and cross-stratified rock units. Qualitative terms describing the character of rock layering are *stratification*, *stratum*, *cross-stratification*, *cross-stratum*, *set*, *coset*, and *composite set*. Quantitative terms applying to the thickness of stratification are *very thick-bedded*, *thick-bedded*, *thin-bedded*, *very thin-bedded*, *laminated*, and *thinly laminated*. Quantitative terms applying to the thickness of units into which the rock splits are *massive*, *blocky*, *slabby*, *flaggy*, *shaly*, *platy*, and *papery*.

A classification of cross-stratification is suggested, based primarily on whether the lower bounding surface of a set of cross-strata is one of erosion or nondeposition and, if erosional, whether it is plane or curved. Features of secondary importance in this classification are the shape of set of cross-strata, the attitude of the axis, the symmetry of the cross-strata with respect to the axis, the arching of the cross-strata, the dip of the cross-strata, and the length of individual cross-strata.

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Introduction

The most common and characteristic structure of sedimentary rocks is stratification or layering. This stratification differs widely in form and size. Many descriptive terms applicable to it have been used in different senses by different authors, resulting in a vagueness, overlapping, and confusion of terms.

A classification and a terminology of stratification are presented that will aid the field geologist in describing differences in primary structures easily and exactly. A review of the historical development of stratification terminology and of the terms and definitions proposed would unduly lengthen this paper. Even a casual reading of current descriptive and reference literature, however, reveals many terms with several definitions.

Geologists should distinguish clearly among three basically different groups of terms: (1) qualitative terms such as stratification, cross-stratification, stratum, and cross-stratum, which are concerned with the attitude and relation of rock layers without regard to scale; (2) quantitative terms such as thick-bedded, thin-bedded, and laminated, which are concerned with the thickness of stratification, or layering; and (3) quantitative terms such as massive, slabby, and flaggy, which are concerned with

the thickness of splitting within stratified units. Thickness of layering and thickness of splitting may differ greatly; terms applying to these two properties should not be used interchangeably.

This report was prepared in collaboration by the University of Arizona and the U. S. Geological Survey and concerns work done on behalf of the U. S. Atomic Energy Commission.

The authors acknowledge the help given in field and office through discussion and through constructive criticism by L. C. Craig, R. P. Fischer, J. W. Harshbarger, C. N. Holmes, T. E. Mullens, C. A. Repenning, G. A. Williams, and others. It is hoped that the classification and terminology set forth in this paper will form a basis for similar discussions by other geologists.

Terminology of Stratification

General Statement

The need for adequate terms for detailed description and comparison of stratification and cross-stratification types is apparent to any geologist who has attempted to analyze these structures. The terms proposed in this paper are based on structures represented in formations that have been examined in the field.

Qualitative Terms

The relations between the qualitative terms presented below are shown in Table 1 and Figure 1.

Stratification is the general term for layering in rocks. It refers both to the process of stratifying and to the state of being stratified, but in this paper stratification is used in the latter sense only. No implication is made concerning the thickness of the individual layers involved. The adjective *stratified* is applicable to any layered sedimentary rock.

A *stratum* is a single layer of homogeneous or gradational lithology, deposited parallel to the original dip of the formation. It is separated from adjacent strata or cross-strata by surfaces of erosion, nondeposition, or abrupt change in character. Stratum is not synonymous with the terms *bed* or *lamination* but includes both. *Bed* and *lamination* carry definite thickness connotations.

Table 1.—Relations among Qualitative* Terms for Stratification

	Horizontally stratified	Cross-stratified
Basic unit; single layer	Stratum	Cross-stratum
Group of strata (or cross-strata) conformable series	Set (of strata)	Set† (of cross-strata)
Group composed of two or more sets	Coset	Coset
Group compounded from strata and cross-strata	Composite set	

* No size connotation is included in these terms.

† Minor sets of cross-strata within larger sets are present in some formations, owing to slight depositional changes, as in degree or direction of dip. No term is here proposed for such variations.

Cross-stratification is the arrangement of layers at one or more angles to the dip of the formation. A *cross-stratified* unit is one with layers deposited at an angle to the original dip of the formation. Many authors have used "cross-bedding" and "cross-lamination" as synonymous with cross-stratification, but it is here proposed to restrict the terms cross-bedding and cross-lamination to a quantitative meaning depending on the thickness of the individual layers, or cross-strata (Table 2).

A *cross-stratum* is a single layer of homogeneous or gradational lithology deposited at an angle to the original dip of the formation and separated from adjacent layers by surfaces of erosion, nondeposition, or abrupt change in character (Fig. 1; Table 1). "Cross-bed" and "cross-lamina" have been used as synonyms of cross-stratum, but it is proposed in this paper that they be restricted to a quantitative meaning. A cross-bed is greater than 1 cm. in thickness, and a cross-lamina is 1 cm. or less (Table 2).

A *set* is a group of essentially conformable strata or cross-strata, separated from other sedimentary units by surfaces of erosion, non-

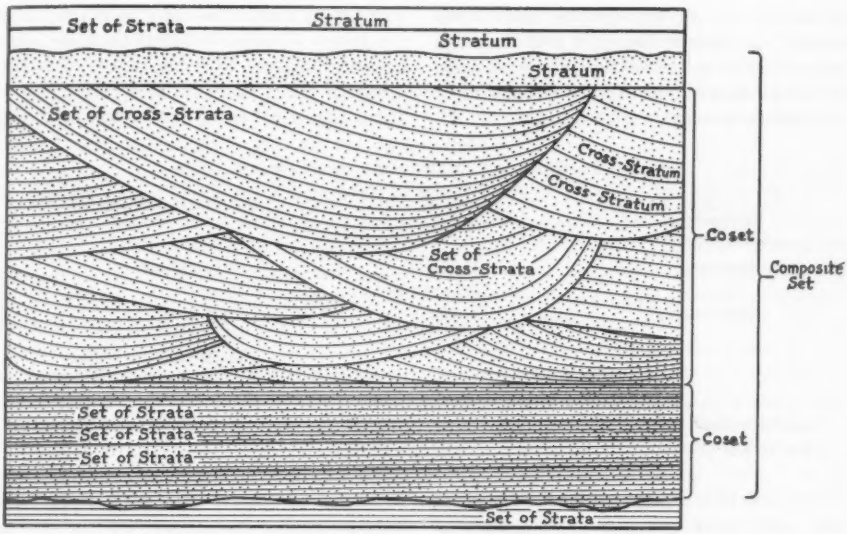

FIGURE 1.—TERMINOLOGY OF STRATIFIED AND CROSS-STRATIFIED UNITS

TABLE 2.—COMPARISON OF QUANTITATIVE TERMS USED IN DESCRIBING LAYERED ROCKS

Terms to describe stratification		Terms to describe cross-stratification		Thickness	Terms to describe splitting property
Very thick-bedded	Beds	Very thickly cross-bedded	Cross-beds	Greater than 120 cm.	Massive
Thick-bedded		Thickly cross-bedded		120 cm. (about 4 ft.) to	Blocky
Thin-bedded		Thinly cross-bedded		60 cm. (about 2 ft.) to	Slabby
Very thin-bedded		Very thinly cross-bedded		5 cm. (about 2 in.) to	Flaggy
Laminated	Laminae	Cross-laminated	Cross-laminae	1 cm. (about ½ in.) to	Shaly (claystone, siltstone) Platy (sandstone, limestone)
Thinly laminated		Thinly cross-laminated		2 mm. (about .08 in.) or less	Papery

deposition, or abrupt change in character (Fig. 1). It is the smallest and most basic *group* unit. Some authors have used the term "cross-bed" in this sense, but, as here proposed, "cross-bed" should apply to a single cross-stratum and should be restricted to a quantitative meaning based on the thickness of the cross-stratum (Table 2).

In some deposits, two or more minor sets (or subsets) of cross-strata occur within a larger, more prominent set. The minor sets may be due to slight variations in direction

or capacity of the transporting agent filling a single scour depression, or to oversteepening and cascading on the lee side of a sand dune. No formal name is here proposed for such minor sets, but their occurrence is recognized.

2 to 4 feet) thick, *thin-bedded* to strata 5 to 60 cm. (about 2 inches to 2 feet) thick, *very thin-bedded* to strata 1 to 5 cm. (about ½ inch to 2 inches) thick, *laminated* to strata 2 mm. to 1 cm. (about .08 to ½ inch) thick, and *thinly laminated* to strata 2 mm. thick or less—that is, paper-thin.

TABLE 3.—GENERAL CLASSIFICATION OF STRATIFIED UNITS

Character of lower surface of stratified unit	Apparent shape of stratified unit	Internal (primary) structure of stratified unit
Erosional	Lenticular	Structureless
	Wedge-shaped	Irregular
Nonerosional	Tabular	Contorted*
	Irregular	Ripple-laminated†
		Horizontally stratified
		Cross-stratified

* See Fairbridge (1946).
† See McKee (1939).

The term *coset* is proposed for a sedimentary unit made up of two or more sets, either of strata or of cross-strata, separated from other strata or cross-strata by original flat surfaces of erosion, nondeposition, or abrupt change in character (Fig. 1).

The term *composite set* is proposed for large sedimentary units compounded from both stratified and cross-stratified units. Thus, a composite set is a sedimentary unit of similar or gradational lithology and consisting of horizontal strata together with cosets of cross-strata (Fig. 1; Table 1).

Quantitative Terms

Bedding and *lamination*, as defined in this paper, denote the thickness of strata. The terms *bed* and *bedding* may be applied to any stratum or stratification of thickness greater than 1 cm. (about ½ inch); *lamina* and *lamination* may be applied to any stratum or stratification 1 cm. or less in thickness. Similarly, *cross-bed* and *cross-bedding* may be applied to any cross-stratum or cross-stratification of thickness greater than 1 cm.; *cross-lamina* and *cross-lamination* may be applied to any cross-stratum and cross-stratification 1 cm. or less in thickness (Table 2).

As adjectives with specific limits it is suggested that *very thick-bedded* be applied to strata greater than 120 cm. (about 4 feet) thick, *thick-bedded* to strata 60 to 120 cm. (about

Bedding and *lamination* should not be confused with terms pertaining to the property of splitting. Terms describing splitting properties are not discussed in this paper, but in Table 2 the quantitative terms proposed for stratification are compared with those that may be used to describe the splitting properties.

CLASSIFICATION OF STRATIFIED UNITS

General Classification

A proposed classification of stratified units is presented in Table 3, which is essentially an outline for describing strata. Except for dimensional and directional measurements, all important elements of stratification are indicated in this table.

The classification is based on three major divisions. The lower surface of any stratified unit (stratum, set, or coset, depending upon the unit in question) is considered basic in the classification. This surface is either *erosional* or *nonerosional*, and it reflects the type of environment immediately preceding deposition.

The apparent shape of the stratification unit is of next importance in classification. It may be *lenticular*, *wedge-shaped*, *tabular*, or *irregular*. These terms are only qualitative, for there are gradations between one shape and another, but the terms adequately describe the chief

varieties. *Irregular* is used especially for such sedimentary masses as bioherms.

Internal structures of the stratified unit constitute a third major division in the classification. They include (1) structureless units due either to depositing agents that do not sort sediment or to destruction of former structures by organisms, (2) irregular strata such as many reef types, (3) contorted beds due to slumping or preconsolidation deformation, (4) ripple lamination, (5) horizontal stratification, and (6) cross-stratification. Most of these internal, primary structures have been thoroughly described and evaluated (Fairbridge, 1946; McKee, 1939); a notable exception is cross-stratification.

Although not included in Table 3, dimensional and directional measurements are part of a complete description of stratified units. Thickness and, in some places, lateral extent of a stratum may be measured. If the stratum is lenticular, it may be practicable to determine the trends of the lines of convergence and of the line of maximum thickness of the stratum. Similarly, it may be practicable to determine the preferred orientation, where present, of internal structures. Measurement of such trends may help define the direction of deposition.

Classification of Cross-Stratification

Detailed studies of cross-stratification are seriously handicapped by the lack of an adequate classification and terminology. Such a classification should (1) be inclusive, (2) make a clear distinction among types of cross-stratification, and (3) be easy to apply to field study.

The descriptive terms most commonly applied to cross-stratification units are "angular" and "tangential." "Torrential" and "regular" have been used as synonyms of "angular." "Angular," "torrential," and "regular" are applied to cross-strata that appear in section as nearly straight lines meeting the underlying surface at high angles. The term "tangential," on the other hand, is applied to cross-strata that in section appear as smooth arcs meeting the underlying surface at low angles. "Angular" cross-stratification, as generally used, implies deposition by water; "tangential" cross-stratification, as generally used, implies deposition by wind. The supposed relationship between genesis and "angular" or "torrential" type is not substantiated by experimental studies nor by observations of structures in modern deposits.

Attempts have been made to classify cross-stratification according to manner of deposition. Such classifications are not all-inclusive, however, because few criteria for determining the mode of formation of cross-stratification are known. For example, cross-stratification of dunes and sand bars may be similar. Perhaps thorough study of cross-stratification in modern sediments and older rocks may develop a feasible genetic classification. At present, however, the most workable classification is a purely descriptive one.

The classification presented here is a modification of one suggested by McKee (1948) and illustrated by Kiersch (1950). The nomenclature is based on physical characteristics rather than on genesis (Table 4; Fig. 2). Pseudo-cross-stratification and variations in horizontal stratification are not included.

The classification is founded on seven criteria, the first of which is basic: (1) the character of the lower bounding surface of the set of cross-strata, (2) the shape of the set of cross-strata, (3) the attitude of the axis of the set of cross-strata, (4) the symmetry of the cross-strata about this axis, (5) the arching of the cross-strata, (6) the dip of the cross-strata, and (7) the length of individual cross-strata.

According to this classification there are three major types of cross-stratification, based on the character of the lower bounding surface of the set of cross-strata (Table 4; Fig. 2): (1) Any set of cross-strata whose lower bounding surface is not a surface of erosion but one of nondeposition or change in character is called a ***simple*** set of cross-strata and is formed by deposition alone. (2) Any set of cross-strata whose lower bounding surface is a planar surface of erosion is called a ***planar*** set of cross strata and results from beveling and subsequent deposition. (3) Any set of cross-strata whose lower bounding surface is a curved surface of erosion is called a ***trough*** set of cross-strata and results from channeling and subsequent deposition. Thus, the basic criterion that determines

TABLE 4.—CLASSIFICATION OF CROSS-STRATIFIED UNITS

Basic Criterion	Subordinate Criteria					
Character of lower boundary surface of set of cross-strata	Shape of sets of cross-strata	Attitude of axis of set of cross-strata	Symmetry of set of cross-strata	Arching of cross-strata	Dip of cross-strata	Length of cross-strata
Nonerosional surfaces (simple cross-stratification)	Lenticular	Plunging	Symmetric	Concave	High angle (> 20 degrees)	Small scale (< 1 foot)
Planar surfaces of erosion (planar cross-stratification)	Tabular	Nonplunging	Asymmetric	Straight		Medium scale (1 to 20 feet)
Curved surfaces of erosion (trough cross-stratification)	Wedge-shaped			Convex	Low angle (< 20 degrees)	Large scale (> 20 feet)

the type of cross-stratification is the character of the lower bounding surface of the set of cross-strata—whether this surface is erosional or not and, if erosional, whether it is plane or curved.

The term "festoon," first applied by Knight (1929) to sandstones of the Casper formation in Wyoming, is a variety of trough cross-stratification. This variety consists essentially of elongate semiellipsoidal troughs that crosscut each other so that only parts of each unit are preserved, and, in section, a festoonlike appearance results. Cross-strata are concave upward and conform to the shapes of the troughs.

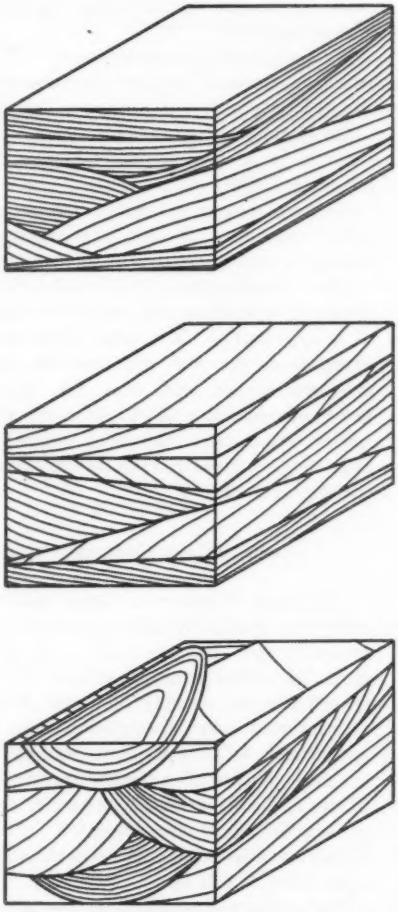
SIMPLE CROSS-STRATIFICATION
The lower bounding surfaces of sets are nonerosional surfaces
PLANAR CROSS-STRATIFICATION
The lower bounding surfaces of sets are planar surfaces of erosion
TROUGH CROSS-STRATIFICATION
The lower bounding surfaces of sets are curved surfaces of erosion

FIGURE 2.—BASIC ELEMENTS OF CLASSIFICATION OF CROSS-STRATIFICATION

Various names other than those proposed in this paper have been used for types of cross-stratification. Some names such as "torrential" and "deltaic" cross-stratification are considered inappropriate because they imply genesis. Others, such as "compound," have been discarded in this paper because the name, although used for the planar type (Lahee, 1941), could apply equally well to the other two basic types and is therefore a source of confusion. If used

at all, "compound" cross-stratification should apply to a combination of the basic types.

Designation of the type of cross-stratification is dependent upon the character of the lower bounding surfaces of sets of cross-strata. In the determination of the three basic types, no restrictions are placed on dimensions, number of imbricating units, convexity or concavity of the filling cross-strata, or agency of formation.

A subordinate criterion considered in the proposed classification is the shape of each set of cross-strata (Table 4). A *lenticular set* is bounded by converging surfaces, at least one of which is curved. A *wedge-shaped set* is bounded by planar, converging surfaces, and a *tabular set* is bounded by planar, essentially parallel surfaces. Nearly all simple cross-stratification sets are lenticular or wedge-shaped, while most planar cross-stratification sets are either tabular or wedge-shaped. All trough cross-stratification sets are lenticular.

A set of cross-strata that is either lenticular or wedge-shaped may be further classified on the basis of its axis—the trace of a line on the lower bounding surface that marks the thickest part of the set as originally deposited. The axis may be characterized as *plunging* or *nonplunging*. Also, the set of cross-strata may be characterized as *symmetric* if the cross-strata correspond in size and shape on opposite sides of the axial plane, or as *asymmetric* if this correspondence is lacking.

Differentiation in some types of cross-strata may also be made by the presence or absence and, if present, by the direction of curvature of individual cross-strata. If the cross-strata arch upward, they are referred to as *convex*; if they arch downward, they are described as *concave*. If, as is rare, the cross-strata are not arched, they are termed *straight*. Concave cross-strata undoubtedly are most common and have long been used as a criterion for telling top from bottom in sedimentary rocks (Shrock, 1948, p. 251–253).

Cross-strata can be differentiated on the basis of degree of inclination. Cross-strata having an average maximum inclination of 20° or more may arbitrarily be considered *high angle*; cross-strata having an average maximum inclination of less than 20° may arbitrarily be considered *low angle*.

For designating the relative magnitude of cross-stratification the following terminology is recommended:

(1) *Small-scale* cross-stratification = cross-strata less than 12 inches (approximately ⅓ meter) in length.

(2) *Medium-scale* cross-stratification = cross-strata 1 to 20 feet (approximately ⅓ to 6 meters) in length.

(3) *Large-scale* cross-stratification = cross-strata more than 20 feet (approximately 6 meters) in length.

This classification of cross-stratification facilitates description and provides a measure of standardization. The classification focuses attention on the most important features of cross-stratification and thus may promote better understanding of the structure. Standardized descriptions may bring out relationships between types of cross-stratification and specific environments.

Summary and Conclusions

This paper is written to define and standardize the qualitative and quantitative terms applicable to stratified and cross-stratified units. Qualitative terms are defined (Table 1). Quantitative terms applicable to the thickness of stratification are distinguished from those applicable to splitting properties, and the limits of these quantitative terms are suggested (Table 2). Terms for describing general features of stratified units are presented (Table 3) Finally, a classification of cross-stratification is presented (Table 4).

The ultimate conclusions of this paper are not the writers', but the reader's. It is for the reader further to test this terminology and classification by application in the field and to determine its value in facilitating description.

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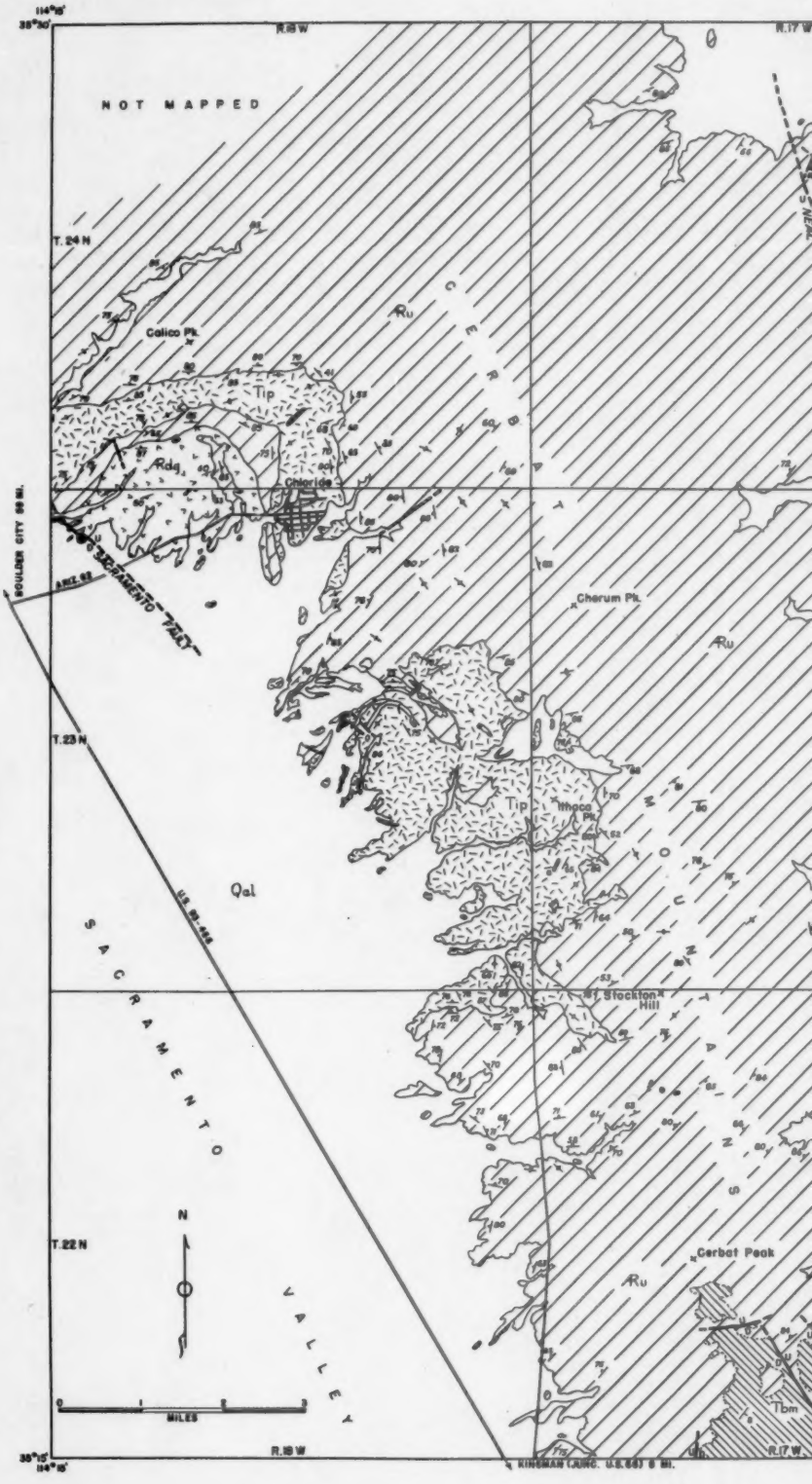
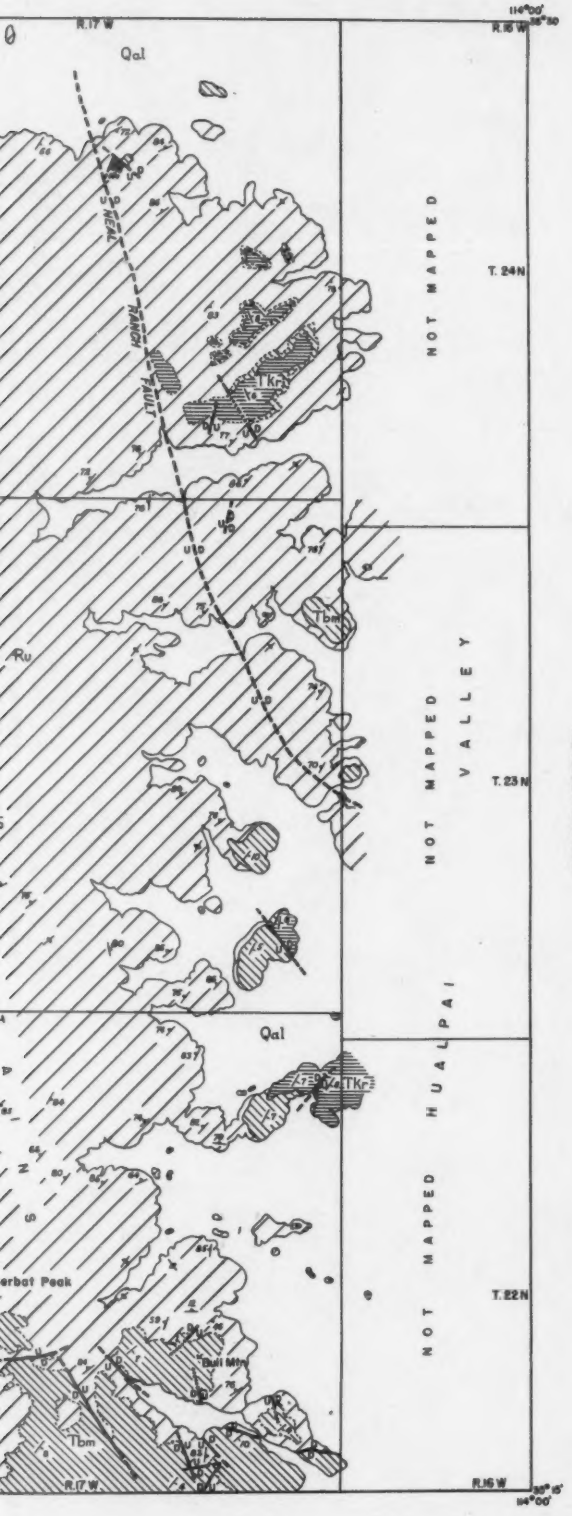


AREAL GEOLOGIC MAP, CHLORIDE QUA

Base map features traced from U. S. Geological Surv
quadrangle, edition of 1944. Geology by Blakemore E. Tho



EXPLANATION

Qal
Alluvium
Bolson deposits, stream terraces, and recent detritus on floors of canyons and washes.

Olivine Basalt
Thin flows, overlying and interrelated with alluvium

QUATERNARY

Tkr
Kingman Rhyolite Series
Flows, tuffs, and agglomerates

Tbm
Bull Mountain Andesitic Series
Flows, tuffs and agglomerates.

TERTIARY

Tip
Ithaca Peak Porphyry
Granite porphyry intrusions

TERTIARY?

CERBAT COMPLEX

Xdg
Diana Granite

Xu
Undifferentiated metamorphic and igneous rocks. Also contains younger dikes.

ARCHEAN

Contact

Approximate Contact

Fault
Where displacement is largely dip slip, U= upthrown side, D= downthrown side.

Probable Fault

Concealed Fault

Strike and Dip of beds

Strike and Dip of foliation

Strike of vertical foliation

LORIDE QUADRANGLE, ARIZONA
eological Survey topographic map of Chloride
kemore E. Thomas. Drafting by Barbara Dunlap.

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 64, PP. 391-420, 6 FIGS. 1 PL. APRIL 1953

GEOLOGY OF THE CHLORIDE QUADRANGLE, ARIZONA

BY BLAKEMORE E. THOMAS

ABSTRACT

The Chloride quadrangle includes the southern two-thirds of the Cerbat Mountains, a range composed mostly of presumably Precambrian crystalline rocks. The oldest rocks are quartzite, mica schist, hornblende-diopside schist, and amphibolite. Younger and larger in amount are granite and granite gneiss. Gabbroic and dioritic gneiss, pegmatite, aplite, diabase, and mylonite also occur as part of the basement complex. The earliest rocks were strongly folded, but igneous intrusion and granitization have obliterated most of the folds. The rock types and the history of the basement complex compare closely with the Archean in the Grand Canyon.

The area contains no Paleozoic rocks and probably no Mesozoic rocks. Intrusions of granite porphyry associated with mineralization are believed to be Tertiary. Dikes of granite porphyry, pegmatite, aplite, and lamprophyre are associated with these intrusions.

Small areas of andesitic and rhyolitic extrusive rocks of Tertiary (?) age occur along the flanks of the range. Quaternary basalt caps these rocks in some places.

Several periods of faulting and erosion are visible in the volcanic rocks. The latest faulting has produced the present mountains. The greatest displacement has been on the west side, and the Cerbat Range is an eastward-tilted fault block, modified in part into a horst. This mountain building seems to have occurred in late Tertiary and Quaternary time. Recent fault-block movement is suggested by the presence of canyon terraces and by the dissection of the pediment at the west base of the range.

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Introduction

Location and Physical Features

The Chloride quadrangle is bounded by meridians 114° and 114° 15′ West, and parallels 35° 15′ and 35° 30′ North and lies within the mountain region of Arizona (Fig. 1) as defined by Ransome (1903). The town of Kingman is approximately 5 miles due south of the quadrangle, Hoover Dam 50 miles to the northwest.

The Cerbat Mountains occupy a central strip, covering approximately two-thirds of the quadrangle. They start immediately north of Kingman and continue beyond the northern border of the quadrangle. The range trends north in the southern half of the quadrangle and northwest in the northern half. The topography is rugged, and some of the slopes are precipitous.

The mountains are flanked on the east by the Hualpai Valley and on the west by the Sacramento Valley, both of which are alluvium-covered basins with gentle gradients sloping away from the mountains. The western side of the Cerbat Mountains presents a relatively uniform face which rises abruptly from the flat skirt of the Sacramento Valley. Numerous canyons and washes, oriented more or less at right angles to the trend of the crest, break the continuity of the face, but a straight line of demarcation still exists between range and basin. On the east side no such line is visible. Instead, a fingering of wide flat washes and mountain spurs occurs, with an abundance of isolated ridges and buttes extending far out into the Hualpai Valley.

A flat pediment, as much as 1½ miles wide, is visible along the west base of the mountains. It s a comparatively even surface, cut without discrimination across all the rock types that compose the mountains, and it has the same general gradient as the surface of the Sacramento Valley. The pediment is veneered with alluvial material, but recent dissection reveals the continuity of the pediment surface. A similar pediment may exist along the eastern mountain base. A few small and scattered bedrock exposures, showing through the alluvium, seem to indicate a pediment. The presence of an extensive erosion surface cannot be proved, however.

Nature and Scope of Work

The field work was conducted in two phases. The first consisted of detailed mapping in the Chloride section of the Wallapai mining district in the summer of 1941. This included location and plotting of all veins. The advance edition of the U.S. Geological Survey's topographic map of the Chloride quadrangle was used as a base map.

The second phase, conducted during the summers of 1947 and 1948, consisted of mapping the areal geology of the balance of the mineralized belt and a section along the east side of the Cerbat Mountains. Much of the mapping was of reconnaissance nature. The U.S. Geological Survey's final topographic sheet of the quadrangle was used as a base map.

The detailed map of the Chloride district and a study of the mineral deposits of the Wallapai mining district have been published (Thomas, 1949). The present paper deals with the general geology of the area.

Work by Others

The first comprehensive geologic reconnaissance to include the Chloride quadrangle was made in 1903–1904 by W. T. Lee (1908). F. C.

Schrader (1909) made a thorough reconnaissance of the mining districts in 1906–1907. His report still serves as a "bible" for the mining operators in the region.

M. G. Dings (1951) made a thorough study of the Wallapai mining district in 1943.

Acknowledgments

This paper has been augmented by the general assistance, suggestions, and constructive criticism of the members of the staffs of the Division of Geological Sciences, California Institute of Technology, and the Department of Geology, University of California, Berkeley. Directly concerned and particularly helpful were Professors C. D. Hulin, Ian Campbell, James A. Noble, and R. H. Jahns.

Drafting of the geologic map was done by Barbara Dunlap.

Terminology

Names for formations follow the usage proposed in a publication on the ore deposits of the Wallapai district (Thomas, 1949).

The color nomenclature in all rock descriptions has been taken from the rock-color chart distributed by the National Research Council (Goddard *et al.*, 1948).

Cerbat Complex

Distribution and Relation to Other Formations

A basement assemblage of metamorphic and igneous rocks forms approximately 90 per cent of that portion of the Cerbat Range included in the Chloride quadrangle. Granite gneiss, of both metamorphic and igneous origin, is the most abundant rock type. Less abundant but widely distributed are discontinuous layers of crystalline schists, small dioritic and gabbroic bodies, highly prophyritic granites, and pegmatite, aplite, and diabase dikes.

The complex is capped along the extreme eastern and southern borders of the range by isolated Tertiary volcanic rocks, most of which have been preserved in down-dropped fault segments. Along the western edge of the Chloride district, most of the basement is terminated by the Sacramento fault (Pl. 1), which has also preserved Tertiary volcanics on its down-dropped side. In the Chloride and Mineral Park districts the complex has been intruded by Ithaca Peak granite porphyry of

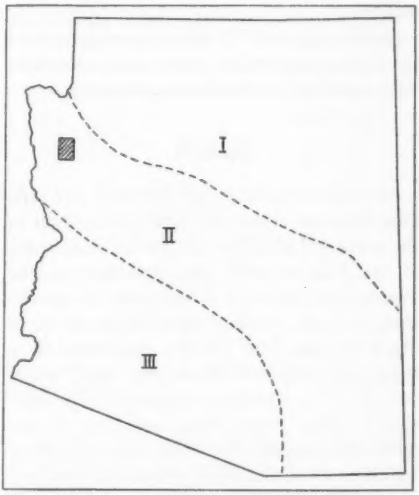

Figure 1. Index Map of Arizona

Showing location of Chloride Quadrangle and Ransome's (1903) physiographic divisions of the state. I—Plateau region; II—Mountain region; III—Desert region.

Tertiary (?) age. Numerous silicic and mafic dikes, along with a host of veins, have likewise invaded these basement rocks.

Metamorphic Rocks—Summary

Vertical to steeply dipping layers of hornblende-diopside schist and amphibolite, biotite schist, quartzite, and *lit-par-lit* gneiss comprise the oldest group of rocks in the region. They may be encountered anywhere in the Cerbat complex. They typically occur together, but in no particular or consistent sequence. Exposures, though numerous, are discontinuous, and more abundant granite gneiss surrounds, interleaves, and grades into the schists. However, in several places enough of these layered rocks have been preserved to reveal their disposition in fragments of large folds. These folds, and evidence of original bedding in the quartzites, suggest that the layered rocks were originally a series of stratified deposits. The initial thickness of the series cannot be estimated, but the

sum of individual layers in the limbs of a large anticline in the Chloride district suggests that it would be measured in thousands of feet.

In addition to the *lit-par-lit* rocks, many of the associated gneisses are believed to be migmatites.

Superimposed on all the major rock types of the Cerbat complex are narrow zones of mylonite, formed by localized intense shearing.

Quartzite

Outcrops of quartzite are scattered and rare. Discontinuous layers 2–3 inches thick occur in the north end of Silver Hill, in the wash north of the Empire mine, and northeast of the Towne vein. A bed 2–3 feet thick is exposed along the north slope of the ridge just north of Merit Spring. One notable occurrence is a steeply dipping layer 40–50 feet thick, which is included in a series of pegmatite and hornblende schist layers lying adjacent to the northern boundary of the Ithaca Peak porphyry intrusion of the Chloride District. An excellent outcrop of this bed is exposed in the northeast corner of Sec. 28, T. 24 N., R. 18 W., just north of the intrusive contact. In all places the quartzite is associated conformably with layers of hornblende schist.

The rock is characteristically light gray, weathering to a moderate brown. It is finely laminated and can be split into thin plates. The texture is interpreted as being blastopsammitic, on the basis of apparently relict subangular to subrounded grains. Recrystallization has resulted in the formation of irregular grains and bands of quartz, most of which are elongated parallel to the layering of the rock. In negligible quantity are dark-brown grains of biotite, wisps of muscovite, irregular grains of magnetite, sericitized grains of andesine feldspar, and stringers of pale chlorite. Tiny needles of apatite are common in much of the quartz.

The quartzite indicates that the oldest series of rocks in the Cerbat complex is, in part, sedimentary in origin. Though some removal and addition of material may have taken place during recrystallization, the almost wholly siliceous composition and the apparently relict fine bedding and blastopsammitic textures are reliable signs of an original layered, fine-grained, quartz sandstone (Bastin, 1909).

Biotite Schist

Biotite schist is widely scattered throughout the Cerbat complex. It occurs both as narrow septa in *lit-par-lit* gneiss and as layers forming separate rock units.

The discrete layers of biotite schist are generally only a few inches to a few feet thick and are commonly interleaved with hornblende-diopside schist and fine-grained granite gneiss. In some places biotite schist seems to be gradational with the hornblende-diopside rocks. The occurrence of small pegmatite or quartz stringers in the schist is not unusual. The well-developed schistosity of the always steeply dipping or vertical strata promotes thorough weathering, and the schist is typically soft and friable.

The schist is dark gray to black where fresh and moderate brown where weathered. Textures are fine-grained lepidoblastic with porphyroblastic phases. A typical specimen contains 50 per cent biotite. The balance consists of oligoclase, orthoclase, microcline, and quartz. Sphene, apatite, and magnetite occur as poikiloblastic inclusions.

Porphyroblasts of oligoclase, orthoclase, and microcline form "eyes" up to 6 mm. in diameter, which are xenoblastic to roughly rectangular, have highly irregular borders, and are intergrown with the adjoining minerals. The feldspars are uniformly clear with unoriented small and large inclusions. The nature of the inclusions (Goodspeed, 1939) plus the clarity and stages of growth represented by small and xenoblastic to larger idioblastic crystals indicate true porphyroblasts and not relict grains or phenocrysts.

The original nature of the schist cannot be determined conclusively. The rock occurs in a crystalline schist series derived partly from sediments and partly from what are believed to have been mafic igneous rocks (now hornblende-diopside schists). The general composition of the biotite schist may represent either sediments or igneous rocks (Harker, 1932, p. 218, 286). Furthermore, the present composition very likely is partly a result of sub-

tractions and additions of material during metamorphism. The layers are narrow and in intrusive igneous surroundings, and a certain amount of metasomatism would be expected.

hypersthene. All gradations exist, from facies containing only hornblende and plagioclase to those with only diopside and plagioclase (Table 1).

TABLE 1.—MINERAL COMPOSITION OF HORNBLENDE-DIOPSIDE SCHIST AND AMPHIBOLITE SUITE*

Hornblende (Color)	Feldspar (% An)	Diopside	Hypersthene	Biotite	Remarks
80 Green	20 (75)				Amphibolite
75 Green	25 —				
55 Brown	45 (55)				
50 Brown	50 (60)			Present	
20 Green	80 (70)			Present	
60 Brown	35 (50)	5			
35 Brown	35 (45)	30			
20 Brown	35 (60)	Present	45	Present	
20 Green	10 (62)	55		12	3% Sphene
	50 (62)	45		5	
70 Brown		10	15	5	Amphibolite

* Prepared with data from 11 thin sections. Figures in per cent.

Hornblende-Diopside Schist and Amphibolite

Somber-hued layers of hornblende-diopside schist make up the bulk of the crystalline schist series and occur throughout the Cerbat complex. They vary from a few inches in width to layers over 400 feet across and in places are several miles long. Thick representative layers are abundant in the lower slopes of the Chloride district.

In fresh specimens the schist ranges from grayish black or greenish black to medium gray. A speckled or banded appearance is common from an alternation of light and dark minerals. The schist characteristically weathers to a moderate or light brown. Most specimens are fine-grained with varying degrees of acicular schistosity, but patches of the rock are massive, either fine- or coarse-grained, and are best termed amphibolites. Many exposures contain narrow stringers of pegmatite, disposed in small folds or in crenulated and contorted ptygmatic patterns. Presumably all of the schists have been intricately folded, but the effects are not revealed by the apparently simple linear schistosity.

Under the microscope the rocks are seen to consist essentially of brown and green varieties of hornblende, calcic plagioclase, diopside, and

The two colors of hornblende are mutually exclusive in any one specimen. Minor minerals are magnetite, apatite, sphene, and biotite.

The rocks that contain contorted stringers pegmatite are similar to those that do not. The only other observable difference is the presence of isolated grains of quartz, orthoclase, and microcline in the usual assemblage of hornblende, plagioclase, and pyroxene.

Rocks of this general nature can be derived from chloritic, calcareous sediments (Harker, 1932, p. 266), but the common source of hornblende schists and amphibolites is mafic igneous rock (Harker, 1932, p. 278–286). Buddington (1939) has described amphibolites with the same mineral assemblage (labradorite, diopside, hypersthene, hornblende) derived from gabbros in the Adirondack complex of northeastern United States. In the Cerbat complex, the hornblende schists occur in narrow, tabular bodies, which undeniably are elements of folds. Therefore, the original rocks are believed to have been basaltic flows or tuffs. Some or all of them may conceivably have been sill-like intrusions in an original series of sediments, but an extrusive origin is in accord with that attributed by Campbell and Maxson (1936) to similar amphibolites in the Grand Canyon, which display relict pillow structures.

Graphite Schist

Graphite schist was observed in two small prospect pits in the west-central portion of Sec. 27, T. 23 N., R. 17 W. These occurrences, though unusual, are interesting evidence for the possible presence of carbonaceous material in the original rocks of the Cerbat complex.

The graphite schist in these exposures varies from 4 to 8 feet in width, strikes northeast, and dips steeply to the east. Any continuation along the strike is concealed by a cover of boulders. The wall rock is granite gneiss. In one pit the graphite schist occupies the east limb of a small faulted anticline.

In hand specimen the schist is medium gray, with a fine-grained lepidoblastic texture. Microscopically it is seen to contain 15–25 per cent graphite, which occurs as abundant microscopic inclusions and as ragged lamellar crystals as much as 0.5 by 2.5 mm. in section. The balance of the rock consists of about 15 per cent orthoclase and microcline feldspar, 30 per cent cordierite, 35 per cent quartz, and less than 1 per cent muscovite. Minor minerals are apatite and sphene.

Migmatite

Lit-par-lit gneiss phase.—Lit-par-lit gneiss, like the crystalline schists, is distributed throughout the Cerbat complex. Outcrops showing distinctive streaked patterns range from a few feet to more than 50 feet wide. The bodies are tabular and may extend longitudinally for hundreds of feet but are discontinuous and typically have indistinct contacts. All are steeply dipping to vertical.

The gneiss is medium gray and weathers to a moderate brown. It has been formed by the injection of sheets of aplitic and pegmatitic granite into bodies of biotite and hornblende schist, producing an alternation of dark and light bands.

The dark bands are composed in part of the minerals previously listed for the schists. In addition, pink almandite (?) garnet is common. In one specimen, from an outcrop just east of the Tennessee mine staff house, an emerald-green garnet (uvarovite?) occurs, as well as almandite. The specimen is further distinctive in that it contains about 3 per cent sillimanite. The sillimanite is present principally as the variety fibrolite, in elongate masses of interlacing needles intergrown with biotite, garnet, quartz, and feldspar. Varying quantities of the invading granitic material are interspersed with the other minerals in the dark bands.

The light bands typically contain 50 per cent or more of quartz and varying amounts of orthoclase, microcline, and sodic plagioclase.

Both light and dark bands range from a fraction of a millimeter to more than 10 cm. wide. The light bands in places pinch and swell irregularly, and the dark bands follow their configuration. Some dark bands are interrupted or fade out in places but continue after an interval. This discontinuity is present both microscopically and in field relations. The contacts of many bands appear sharp in hand specimen, though the border minerals always interlock, whereas some contacts are hazy and indistinct. All degrees of transition exist between aplite or pegmatite bands and schist, and thorough mixing has produced many narrow layers that are indistinguishable from the granite gneiss that forms the bulk of the Cerbat complex. The field boundaries of the exposures of *lit-par-lit* gneiss are of similar transitional nature.

Ptygmatic folds, always present in the pegmatite and aplite sheets, show the usual characters of extreme contortion, lack of correspondence in form with the fabric of the enclosing schist, lack of competence, and absence of correlation between forms assumed by different layers in a single outcrop or specimen. Difference in form is well demonstrated where a contorted layer is adjacent to relatively straight, undeformed sheets. There seems to be a uniform trend to the axial planes of the folds, roughly parallel to the foliation of the host rock.

All observed foliation of the *lit-par-lit* gneiss is conformable with the structural grain of the region. Nothing indicates that the attitude of the invaded schists has been disturbed.

All these features are strikingly similar to those described by Fenner (1914) for Precambrian injection gneisses in northwestern New Jersey. Fenner's analysis of the mode of origin of these gneisses is believed to apply equally well in the Chloride quadrangle. This mode is one of active invasion by a highly fluid, silicic magma. The magma quietly entered the layers

of a foliated rock, causing no violent disturbance of the attitude of the folia. The host rock was impregnated and somewhat softened by an "advance dilute portion" of the magma, which facilitated the injection process and initiated transformation of much of the invaded material.

Other composite gneiss phases.—A detailed consideration of the possibility that much of the gneiss may be migmatitic instead of magmatic in origin will not be attempted. In the *lit-par-lit* gneisses, all degrees of transition from schist bands to granite gneiss occur. The classification of the latter as a composite rock, formed by processes of metasomatism and magmatic injection, is valid, at least for the areas where such gradational relationships exist. Whether extrapolation may be applied, and a similar mode of origin assumed for gneisses that look the same but lack exposures showing gradation, is largely a matter of opinion. It is believed that much of the gneiss, perhaps 50 per cent or more, is migmatite, and the balance of magmatic origin.

Evidence of at least two ages of gneiss supports this view. For example, along the crest of the range in the Summit mine area, thin septa and lenses of fine-grained gray gneiss, transitional with layers of hornblende schist, are surrounded and cut by a later medium-grained gneiss. The sharp transgressional boundaries of the latter are taken to indicate that it is an igneous or orthogneiss, whereas the former is apparently a migmatite derived from hornblende schist. Repetitions of these relationships occur throughout the range.

A suggestion that migmatite is present in large amounts is presented by the numerous areas where granite gneiss is transitional with layers of schist and *lit-par-lit* gneiss. This is illustrated in the high mountain slopes east of Chloride, where granite gneiss composes at least 90 per cent of the rocks yet rarely continues for more than a few hundred feet without grading into discontinuous septa of biotite and hornblende schist or patches of *lit-par-lit* gneiss.

The presence of ptygmatic folding in areas of granite gneiss is thought to indicate migmatization. In some places highly contorted pegmatite veinlets exist in what otherwise appears to be a simple granite gneiss. The walls of the veinlets are distinct, but the border minerals interlock with the minerals of the gneiss. The axial planes of the folds are roughly parallel with the foliation of the rock.

Further support for widespread migmatization is the fact that throughout the Cerbat complex the foliation in the gneisses is conformable with the textural and structural grain of associated schists. Undoubtedly some of this foliation has been inherited from previous schists. However, emplacement of intrusive rocks has been influenced by pre-existing structure, and any flow banding should be conformable to this structure. Furthermore, uniform directional stress apparently existed before, during, and after intrusion of most of the igneous rocks in the complex, and foliation produced by it would likewise parallel the regional trend.

Granitization, in the sense that it is synonymous with migmatization (Turner, 1948, p. 306), has seemingly been of considerable importance in the formation of a large portion of the Cerbat complex. However, even with extremely detailed study, precise resolution of areas of granitization and areas of purely igneous gneiss would be difficult, if not impossible.

Mylonite

"In view of the often complex tectonic operations with which regional metamorphism is closely connected, it cannot be supposed that shearing stress merely suffers, like temperature, a steady decline. Rather must we expect incidental revivals of more or less intense shearing stress, recurring, it may be, at various stages of the continued fall of temperature."

This generalization by Harker (1932, p. 346–347) is well borne out in the Cerbat complex, where effects of continued stress are visible in the microstructures of both metamorphic and igneous rocks. Intense local shearing is evidenced by zones of mylonitic rocks at many widely scattered points in the complex. Examples are exposed just southwest of the Arizona Magma dump, in the Boulder Dam claim, at the Towne mine, south of Mullen ranch, and in the saddle of the ridge in the extreme southeastern portion of the quadrangle. In all places the mylonite has a northeast trend, parallel with the general structural grain. The zones range from a few feet to 10–12 feet wide. They

have no distinct borders but pass gradually into the surrounding gneissose or schistose rocks.

In hand specimen the mylonites are greenish gray to dark gray and appear much like a porphyroblastic slate or phyllite. They are either massive or finely laminated. In some places slickensided surfaces exist. Augen of feldspar are common, ranging in size from barely visible specks to lenticular crystals over an inch long. The larger grains were observed where pegmatite veinlets seemed to be part of the parent rock.

Under the microscope the effects of intense shearing and crushing are reflected in dark laminae composed of mineral aggregates so fine-grained that they are sometimes not resolvable under highest magnification. Where identification is possible, the minerals are principally sericite and chlorite, with some epidote, quartz, and feldspar (probably albite). Interleaved with the dark bands are aggregates, lenticles, and stringers of quartz, typically intercrystallized with small grains of orthoclase. Sinuous, pinching and swelling bands of quartz indicate flowage and recrystallization. Augen of orthoclase are numerous, and augen of microcline occur in some specimens. All the augen seem to be relict grains and are affected by cataclasis but not by recrystallization. Other relict minerals observed are tiny grains of almandite garnet, scattered sparsely along one or two laminae of a specimen, occasional flakes of biotite, and a few crystals of sphene. These are anomalous minerals, high-grade metamorphic products in rocks which have suffered extreme cataclasis with accompanying recrystallization to the low-grade mineral assemblage sericite-chlorite-epidote. On the basis of their retrogressive metamorphic character, the rocks may appropriately be classed as phyllonites (phyllite mylonites).

The localized intense shearing and the resultant zones of mylonite are assigned to the Precambrian, essentially because wide mylonite zones were observed only in the areas of the Cerbat complex. Owing to a gap in the geologic record, involving probably all the Paleozoic and Mesozoic, this tenuous assignment is the best that can be made.

Conditions of Metamorphism

The metamorphic rocks of the Cerbat complex have evolved through the action of several metamorphic processes. The chief causes have been orogenic forces and plutonic igneous intrusion.

The first major phase of metamorphism was produced by orogenesis, with strong directed pressure and probably much heat of geothermal and mechanical origin. Originally layered rocks were squeezed into large folds and recrystallized to form a series of crystalline schists. These dynamothermal effects encompass and extend beyond the area studied, and the action may be termed "regional" metamorphism. The essential characteristics of the crystalline schists are believed to stem from these early effects.

A second phase of intensive metamorphism was imposed through igneous intrusion, which appears to have followed closely or to have been concomitant with the last stages of orogenesis. The intrusion was characterized by injection phenomena and the production of migmatite. The metamorphism attributable to it is classed as "plutonic." Following Turner's definition (1948, p. 9), this means essentially deep-seated regional metamorphism, at high temperatures and pressures, accompanied by deformation and characterized by metasomatism, infiltration, and injection phenomena derived from igneous intrusion.

Using the concept of metamorphic facies (Turner, 1948, p. 54), the grade of metamorphism that evolved under these combined dynamothermal and plutonic conditions can be determined. The hornblende-diopside schists and amphibolites provide the critical sensitive mineral assemblage needed to indicate the grade. The association of hornblende and calcic plagioclase embraces both medium- and high-grade regional metamorphism and corresponds in general to the sillimanite, kyanite, staurolite, and, in part, the garnet zones of the Scottish Highlands (Turner, 1948, p. 76). The occurrence together of plagioclase-hornblende-diopside-hypersthene represents a special high-grade subfacies which not only can be assigned to the amphibolite facies, but also has characteristics of the granulite facies. It is analogous to a mineral assemblage in the Archean rocks of the Adirondack complex, where intense regional

metamorphism has been accompanied and aided by injection of granitic batholiths (Turner, 1948, p. 104). It also corresponds to the sillimanite zone of regional metamorphism. In connection with this, the *lit-par-lit* gneiss that contains the assemblage sillimanite-almandite-orthoclase is noteworthy.

No conclusions can be drawn as to the evolution of the migmatites. Mechanical injection of pegmatitic and aplitic material is indicated in the *lit-par-lit* forms, and magmatic soaking and impregnation of the host rocks probably preceded and accompanied injection. The source of the pegmatite and aplite is obscure. Igneous origin is possible, but production by differential fusion (anatexis) is perhaps more likely, because the pegmatite and aplite sheets cannot be correlated as differentiates of any adjacent large intrusion. The anomalous appearance of pegmatite and aplite, as the first plutonic products in the Cerbat complex, is more easily explained as a function of differential fusion than of late differentiation of a hidden batholith, considering the presumably great depth of erosion of Archean terrains.

Deformation appears to have continued from the early phases of metamorphism, through the plutonic stages, and even to have outlasted the crystallization of the igneous rocks, thereby resulting in retrograde metamorphism. Continued stress during the injection of magma is suggested by the ptygmatic folding in the migmatites. Correspondence between the outlines of the folded igneous layers and the trend of schistosity in the host rock implies injection under conditions of active deformation and signifies that orogenic folding was not yet complete at the time of first magmatic intrusion.

RETROGRADE METAMORPHISM: In regionally metamorphosed rocks retrogressive changes may be brought about by declining temperatures and by variations in stress, producing both mineralogical and structural modifications. Both types of change have affected the metamorphic rocks of the Cerbat complex, but the effects of stress far outweigh those of temperature.

Ignoring near-surface changes due to weathering and restricted alteration caused by hydrothermal sulfide mineralization, only a few mineralogical changes can be attributed solely to declining temperature at depth: small amounts of chlorite developed from garnet and biotite, amphibole from diopside, biotite from hornblende, distinct crystals of zoisite and epidote from plagioclase (rather than hydrothermally produced obscure grains in saussurite), possibly some sericite from potash feldspar, and microperthite in orthoclase and microcline. The exsolution of albite is the most common mineral change.

Structural changes are much more in evidence. Continuation and perhaps intensification of shearing stress during the gradual decline of temperature has left its widespread mark. Characteristic features are fractures and displacement in garnet, pyroxene, and hornblende, granulation and crushing of the component grains in the rocks, especially quartz and feldspar, undulatory extinction and deformation of twin lamellae, the presence of sheared "augen" in many gneisses, and extreme comminution, stretching, and milling in the mylonites. The mylonites are the best examples of well-developed retrogressive metamorphism; in them intense deformation has been accompanied by mineralogical reconstitution from a high-grade assemblage to one containing the low-grade minerals chlorite, epidote, sericite, and albite.

Granite Gneiss

All the granite gneiss in the region is discussed under this heading. However, granitization or migmatization is believed to have been widespread, and many of these rocks may actually be composite in origin.

Granite gneiss is the principal constituent of the Cerbat complex and the most abundant and widespread rock type in the quadrangle. It forms massive rugged surfaces with individual erosional shapes tending toward spheroids or ellipsoids. Where jointing is strong, sheer-walled blocks and columnar crags have developed.

The rocks are characterized by diverse texture and structure but relatively uniform composition. Exposures with no obvious gneissic banding exist, but the majority show every range from faintly banded types to flaser gneisses with well-developed "augen" of feldspar. Aside from phenocrysts (or porphyroblasts or porphyroclasts) which reach more than 3 cm. in length, the rocks are predomi-

nantly fine-grained (less than 3 mm.). Most specimens are light to medium gray when fresh, but some are pinkish gray and others light bluish gray. Weathered surfaces are ordinarily light brown to moderate yellowish brown. Segregations of biotite as much as 6 mm. in diameter give a speckled appearance to many specimens.

Under the microscope, the diversity of texture and structure is even more marked. Most of the gneiss has been cataclastically deformed, and the component minerals are surrounded by a microcrystalline mosaic of crushed grains. However, some specimens have strong gneissic banding but show no crushing or granulation.

The principal minerals are orthoclase (50–70 per cent) and quartz (10–20 per cent). In most specimens microcline and oligoclase are present in small amounts. The only characterizing mafic mineral which appears consistently is biotite. A few grains to a few per cent of green hornblende and some sodic andesine were observed in some darker granite gneisses.

Minor accessory minerals are scattered grains of apatite, sphene, magnetite, and zircon.

It has been emphasized that the gneisses may be of either composite or igneous origin. If the two types are present in about equal proportions, their widespread occurrence indicates that the igneous members represent an intrusion of batholithic proportions, which has engulfed, assimilated, displaced, and migmatized large quantities of schist.

The gneissic structures are of complex derivation. Some of the foliation has been inherited from schists, some probably represents flowage controlled by structures of the intruded rocks, and some may be due to crystallization under directional stress. Superimposed on these is cataclastic deformation, which has emphasized the foliation at times and in places may have been the sole agent. This later shearing, while localized in intensity, paralleled the established structural trend and did not produce any noticeable discordant features.

Mafic Gneisses

No large exposures of mafic gneiss occur, but examples are abundant throughout the areas of the granite gneisses. Most of these small isolated bodies are probably composite rocks, but some might be mafic segregations of magma. Several distinctly tabular bodies could represent either migmatized schists or intrusive sills.

The rocks are medium to dark gray and weather to a moderate brown. They are fine- to medium-grained, may be granular massive, but usually show faint to strong gneissic banding. Tonalite and quartz gabbro were determined under the microscope.

Diana Granite

The Diana granite, a unit of the Cerbat complex, is a coarsely porphyritic intrusion, which was mapped separately in the Chloride district. Other porphyritic granites occur elsewhere, notably along the east side of the range. They are similar in general appearance and might be mapped as correlatives if detailed work were applied to the entire quadrangle.

This coarsely porphyritic body occupies the central part of the western half of the Chloride district. It is roughly circular, with a diameter of about 1¾ miles.

The medium to coarse-grained texture of the rock makes it particularly susceptible to weathering and erosion. This, and its location along the base of the mountain range, has contributed substantially to the development of a pediment. The effect has been so marked that the pediment reflects the distribution of the rock, forming a reentrant into the mountains which outlines the dimensions of the body. The surface is essentially flat, sloping gently southwest, and it blends insensibly with the overlapping alluvium. The pediment is subject to interruptions only where isolated septa or pendants of more resistant older schist and gneiss occur. A few more resistant portions of the Diana granite have yielded spheroidal boulders, but these are not common.

The emplacement of the Diana granite was controlled by the distribution of the crystalline schists. The granite fills the core of a large northeast-pitching anticline and is virtually surrounded by steeply dipping layers of schist, gneiss, and pegmatite. Broadly, and many places in detail, its contacts with the schist and older gneiss are distinct. However, in

places interaction and apparent assimilation have produced narrow zones of composite rocks. Some border schists show intimate sheet injection that seems to stem from the Diana granite.

One thick septum or pendant of schist and gneiss, 1½ miles long and three-sixteenths of a mile wide, occurs in the eastern portion of the intrusion and forms a distinctive ridge in the otherwise uniform surface. Other discontinuous bodies of older rock are scattered through the granite, but the largest of these is only 200 by 800 feet at the surface.

The Diana granite is cut by numerous mafic and silicic dikes. Along part of its western border it is in fault contact with a series of Tertiary extrusive volcanic rocks.

The rock is medium gray on fresh fracture and weathers light or moderate brown. Many of the outcrops show weak foliation, and some exposures are highly gneissic or mylonitic. The texture is porphyritic, with a medium-grained groundmass.

Tabular, slightly embayed orthoclase phenocrysts, as much as 5 cm. long, make up about 20 per cent of the rock. The groundmass, with an average grain size of 2 mm., consists of about 40 per cent orthoclase, 30 per cent microcline, 10 per cent oligoclase, 15 per cent quartz, and 5 per cent biotite. Apatite, sphene, and zircon are additional accessories.

A plutonic igneous origin for the Diana granite is suggested by (1) its relative purity, contrasted with the heterogeneous and thoroughly mixed areas of schist and older granite gneiss; (2) its control by pre-existing structure, with limitation to the core of an anticline; (3) its igneous texture, particularly the large phenocrysts which, though poikilitic, do not contain wholesale inclusions of all the other minerals, as in the poikiloblastic "sieve" pattern so characteristic of metamorphic rocks.

The strongly gneissic phases are not abundant and apparently formed in response to the same localized post-consolidation shearing stresses imposed on the rest of the Cerbat complex. Mylonites in the Diana granite have the same northeast trend as those elsewhere. Gneissic structures at and parallel to the borders of the intrusion are classed as flow banding.

Emplacement of the Diana granite seems to be later than the formation of much of the Cerbat granite gneiss, for it cuts across bands of granite gneiss that are interleaved with schists. Its inclusion in the Precambrian is warranted only by the lack of any better determination.

Pegmatite

In addition to the pegmatite layers in the migmatites, dikes of all sizes are numerous in the Cerbat complex. These range up to masses as much as 600 feet wide and a mile or more long, and their distinctive outcrops are often discernible for several miles. Large white pegmatite bodies are particularly well displayed in the limbs of the large anticline in the Chloride district. Here they were injected as sill-like masses between layers of dark schist and stand out in sharp color contrast. Some pegmatites form irregular lenses and pockets.

The pegmatites are granitic. Gray to milky quartz composes 30 to 40 per cent of the rock; the remainder is orthoclase and minor amounts of light bluish-gray microcline. Departures from a prevailingly simple composition were noted in only a few pegmatites. In a specimen from the area just east of the Alice claim all the feldspar is oligoclase, strongly antiperthitic with stringers and patches of orthoclase. A dike northwest of the Dorothy prospect contains a small amount of black tourmaline in a graphic intergrowth with gray quartz. Muscovite was observed in several dikes but is rare. Tiny rutile crystals occur in some of the muscovite. A small quantity of beryl was found in pegmatite on the claim at the south end of the Jim Kane group (Haury, 1947). At the open-pit mine of the Consolidated Feldspar Corporation, on the southeast slope of Bull Mountain, large masses of quartz and orthoclase contain small seams and segregations of biotite and rough prismatic crystals, a few inches long, of black vitreous-appearing gadolinite. Detailed study of this deposit might reveal the presence of other rare minerals.

At least two ages of pegmatite are present in the Cerbat complex. The earlier, sheets and veinlets in the *lit-par-lit* gneisses, is interpreted as the oldest intrusive rock of the complex and is possibly a product of differential fusion (anatexis). Many of the larger intrusions of pegmatite may also be of this age.

Younger pegmatites cut the Diana granite

and similar porphyritic granite bodies in the complex. Their post-Diana age suggests an origin as late products of magmatic differentiation. The dikes that contain tourmaline, muscovite, beryl, and gadolinite, which require volatile "mineralizers" in their formation, probably belong to this later group. Other dikes, which lack such minerals and cannot be correlated with the Diana granite, may belong to either age as far as composition and appearance are concerned.

Interpretation of the Precambrian pegmatites is complicated by the fact that some of the dikes in the complex are probably post-Cambrian, for the later Ithaca Peak granite porphyry intrusion was followed by pegmatites similar to these simple granitic types.

Aplite

Aplitic phases occur in the pegmatites, and isolated small aplite dikes occur in some places. The section on origin and age of the pegmatites is equally applicable to the aplites. The composition is similar to that of the simple pegmatites except that microcline and orthoclase are more abundant, quartz less abundant, and many small dikes carry 1 to 2 per cent of almandite (?) garnet.

Diabase Dikes

Fine- to medium-grained diabase dikes cut all the other rock types of the Cerbat complex. They range from a few feet to as much as 150 feet wide, and some can be traced for more than a mile. Excellent examples can be seen at the Hercules and Midnight mines and in the area just north of the Arizona Magma mine. Vertical to steeply dipping attitudes are typical, but three dikes that cut the ridge east of the Mary Bell mine strike N. 40–50° W. and dip only 15°–40° E. Diabase occurs in one or both walls of many of the quartz-sulfide veins.

Fresh specimens are dark gray; weathered forms are light brownish gray to moderate brown. Most of the dikes have been thoroughly altered either by weathering or by hydrothermal action. The dikes range from very fine-grained chilled border facies to examples in which the average length of the feldspar laths exceeds 3 mm.

A typical specimen contains approximately 65 per cent labradorite (An_{60}), 25–30 per cent pigeonite, 1–2 per cent hypersthene, about 5 per cent magnetite, and small amounts of biotite and apatite.

In several places the diabase dikes invaded fracture lines previously occupied by Precambrian pegmatites. Appreciable digestion of the pegmatite resulted in a pink-tinted hybrid rock. Clumps of partially consumed pegmatite, grading into the general mixture, identify the mode of formation.

Since diabase was not observed to intrude any rocks but those of the Cerbat complex, it is assigned to the Precambrian. However, as the youngest intrusive rock of the complex, it could be of almost any age. It possibly came in at the same time as the diabases of the Grand Canyon, in which event it would be Algonkian (L. F. Noble, 1914, p. 55–60).

Archean History of the Cerbat Complex

The Archean history of the Cerbat complex can be divided into four broad stages:

(1) The first decipherable event in the region was extensive volcanism, with the accumulation of probably thousands of feet of basaltic rocks. At intervals throughout this period quartz sandstones and shales were deposited, possibly grading in places into volcanic material. The amount of these sediments seems to have been small, but granitization and igneous intrusion have masked a large volume of the older rocks, much of which was perhaps sedimentary.

(2) An intense orogenic disturbance (Arizonan revolution, Hinds, 1935) followed the accumulation of the volcanic and sedimentary strata. The beds were folded and subjected to high-grade regional metamorphism, with horizontal compression imposing a uniform northeast structural trend on the region. Probably near the end of the orogeny, pegmatitic solutions invaded the deformed rocks. *Lit-par-lit* injection took place on a large scale, and all the rocks apparently became soft and plastic, for ptygmatic folding was widespread. Much migmatization of the schists occurred, obscuring parts or all of many folds. However, the resulting granite-gneiss migmatite inherited the structural grain of the schists, and the foliation in general conforms to the regional trend. In

places the foliation clearly follows the course of plunging folds. Fundamentally unchanged directional stress continued through and beyond this stage.

(3) Intrusion of granitic magma, possibly batholithic, followed this preliminary "soaking" and sheet injection phase. The emplacement of the granite bodies was influenced by the structure of the schists, and much primary gneissic banding, parallel to the guiding schist layers, is believed to exist. What process was most important in the intrusion of the magma is not known. Some displacement of the preexisting rocks probably took place, and the occurrence of xenoliths suggests stoping. Assimilation products are well developed at the contacts of bodies such as the Diana granite. Whatever the process, the result has been the obliteration of continuous folds, leaving only scattered remnants. The introduction of large quantities of pegmatite, with auxiliary aplite, marks the close of this intrusive stage.

Continuation or revival of stress, with unchanged orientation, impressed secondary foliation or heightened primary gneissic structure in the bulk of the igneous intrusions. Localized intense shearing, after consolidation of the plutonic bodies, formed narrow zones of mylonite throughout the Cerbat complex.

(4) The final event was deep erosion and peneplanation. This may have taken place in late Archean time. However, the preserved portions of the surface lie beneath Tertiary volcanic strata, and erosion during intervening ages could have formed or perfected this surface.

The diabase dikes of the Cerbat complex are believed to be much later, though possibly still Precambrian (Algonkian).

Age and Correlation of Cerbat Complex

The lack of fossiliferous rocks in the Cerbat Range precludes any conclusive age determinations, and the Cerbat complex is placed in the Archean on the basis of petrologic similarity to other Archean terrains. Its "Archean characteristics" are: a high grade of regional and plutonic metamorphism; extensive injection of magmatic material under conditions of general softening and plasticity; and the dominance of true granite in the intrusive members.

The most logical and accurate correlation is with the Archean rocks of the Grand Canyon, which possess similar lithological, structural, and historical features. Campbell and Maxson (1938) have described four periods in the Archean history of the Grand Canyon: (1) deposition of thick, monotonous sedimentary formations, dominantly sandy clays with quartz sands; (2) eruption of basalts, with intercalated thin sediments; (3) orogeny and granitic intrusion; and (4) erosion. Except for the lack of thick sediments (which may have been granitized) in the Cerbat complex, this sequence of events and the resulting assemblage of quartzites, mica schists, hornblende schists, amphibolites, many migmatites, and large volumes of granite and pegmatite are almost identical in the two regions. Application of orogenic forces was also the same, resulting in a northeast structural trend. Campbell and Maxson suggested restricting the name Vishnu series to the metasediments, and in that sense it has no correlative in the Cerbat complex. In the broader sense, as used by L. F. Noble (1914, p. 32–35), however, the Vishnu schist can be correlated with the Cerbat schists.

Lithologically and structurally the Cerbat complex is similar to the Archean rocks in the core of the Virgin Range, Nevada, briefly described by Longwell (1928, p. 22). With other areas, correlation is weak and can be based only on the northeast structural trend. Wilson's (1940) summary of the Precambrian of Arizona basin ranges shows this general northeast grain to be characteristic wherever the early Precambrian has been described in Arizona, including the Pinal and Yavapai schists of the south and central areas.

Ithaca Peak Porphyry

Distribution and Topographic Expression

Two intrusions of granite porphyry, within the Cerbat complex but presumably much younger, are exposed along the western flank of the range, one in the Chloride district and the other in the Mineral Park district. They are similar in texture and composition and are assumed to be the same age.

The intrusion in the Chloride district is in the form of a curved, steeply dipping sill 2000–5000 feet thick. North of Chloride the granite porphyry has been eroded into a group of rugged ridges and peaks that rise to heights of 1000 feet above the town. The remainder of the arc, to the west, consists essentially of one main ridge that is much less rugged and which, aside from isolated peaks, decreases gradually in elevation from east to west.

The intrusion in the Mineral Park district is apparently a roughly equidimensional stock, 3½–4 miles in diameter, with its center approximately at Gross ranch. The intrusion is flanked on the west by alluvium and continues an unknown distance beneath the surface of the Sacramento Valley. Several thick tongues extend from its southeast side, the largest reaching 1½ miles through Union Basin to the Golconda mine. A gently rolling pediment, with exposed widths up to a mile, has been developed along the western side. The portion of the stock within the mountains has been eroded into a cluster of steep rugged peaks with an average relief of about 700 feet. These stand out in sharp contrast to the surrounding areas of the grayish Cerbat complex, as they have weathered to light brown and reddish brown, and also have rougher slopes. Most of the rough slopes and crags are due to silicification of the rock.

In addition to this large sill and stock, numerous dikes or sills and a few small plugs of granite porphyry have invaded the Cerbat complex. Dikes are especially abundant in the Chloride district.

Petrography

Fresh specimens are light gray. Surface exposures weather to light brown or pale reddish brown and in parts of the Mineral Park stock to dark reds and brown. Much of the stock has been mineralized, and the oxidation of pyrite accounts for the reddish color.

Parts of both intrusions are foliated, and some border facies are strongly gneissic. Foliation is better developed, however, in the Chloride district sill than in the Mineral Park stock.

The rock is hypidiomorphic porphyritic, with a fine-grained to microcrystalline groundmass. Microcrystalline, interlocking grains of quartz and feldspar compose 10–15 per cent of the rock. An average specimen contains about 60 per cent orthoclase, small amounts of microcline and oligoclase, 30 per cent quartz, and 5–10 per cent biotite. Accessory minerals include magnetite, sphene, zircon, apatite, and, in some specimens, garnet.

Origin and Intrusive Relations

The microcrystalline phase of the groundmass in the Ithaca Peak porphyry suggests a relatively shallow, hypabyssal intrusive. This is its chief distinguishing characteristic; in composition and even in gneissic structure it is highly similar to much of the Precambrian granite.

The body in the Chloride district was intruded concordantly into curved layers of Archean schist and gneiss, and its exposure emphasizes the banding of the outcrops that surround a core of Diana granite. It is classed best as a curved sill that was injected in a near-vertical attitude into a steeply pitching large anticline.

The Mineral Park stock was likewise influenced by pre-existing structure, though not to the same degree. Its northwestern edge follows a steeply pitching anticline, and anticlinal septa of Archean rocks are present in the northern portion of the stock. The configuration of its southern border and the attitudes in the adjacent Cerbat complex suggest that it has also followed an anticline there. In many places, however, its contacts parallel the foliation of the adjacent country rock but do not follow distinct folds. In a few places it is strongly discordant with neighboring structure.

The contacts of the intrusions are sharp, though the schistosity of surrounding Archean rocks has in places led to narrow zones of closely spaced injected sheets of granite porphyry. The high-grade regional metamorphic and plutonic igneous nature of the host rocks has prevented the development of any obvious contact metamorphic effects.

The Mineral Park stock is believed to have been emplaced by a combination of forceful injection and piecemeal stoping. The structural guidance of folds and the planar-flow structures at and near many of the contacts uphold the

former. The presence of isolated pendants of the Cerbat complex in the interior of the stock supports the latter. In the Chloride district sill some stoping may have taken place, though spreading of the anticlinal layers conceivably could have provided the necessary space.

Age

Schrader (1909, p. 30) classified the Ithaca Peak porphyry as late Jurassic or early Cretaceous and thought it to be of the same period of intrusion as the batholiths of California and western Nevada. However, he also believed the intrusions to be allied with similar bodies in the Oatman district, 25 miles southwest of Mineral Park. These were later mapped as intrusive into volcanic strata and assigned to the Tertiary (Lausen, 1931, p. 44–46). A Tertiary age is assumed for the Ithaca Peak porphyry in this paper, on the basis of its close areal and structural relations with the Tertiary vein system in the Cerbat Range.

Granite Porphyry Dikes

Vertical, or nearly vertical, tabular sheets of granite porphyry are scattered throughout the areas adjacent to the Ithaca Peak porphyry intrusions. They are profuse in the eastern part of the Chloride district. Although many of the sheets are concordant with the schistosity of the host rocks and should properly be called sills, they are classed as dikes for this discussion.

The dikes range from discontinuous sheets a few inches thick to layers as much as 200 feet thick and several miles long. Though many of the prominent dikes strike east-northeast, the major trend is northwest. Granite porphyry is common in one or both walls of the quartz-sulfide veins.

Some of the dikes form ridges, but most have no distinctive topographic expression. All the dikes seem to have undergone some hydrothermal alteration and silicification, even where they are not in direct association with quartz veins. As a result, the dikes are bleached whitish to yellowish brown, which stands out against the grays and blacks of the surrounding gneisses and schists.

A porphyritic texture in the dikes is provided by relatively few orthoclase phenocrysts, which, however, occur as crystals as much as several millimeters in length. A common textural variation is provided by the occurrence of small stringers, lenses, and pods of white pegmatite, apparently younger but closely related genetically to the granite porphyry. The bulk of the rock is fine-grained and aplitic in hand specimen. Brown limonite-stained joint lines and faces are typical, with strong staining extending from the fractures into the rock as much as 1½ cm. Many of these fractures are lined with chalcedonic quartz.

The microscope shows the groundmass to be an allotriomorphic aggregate of quartz, microperthitic orthoclase, and a few grains of microcline. A few small aggregates of chlorite, quartz, magnetite, and limonite apparently have been formed from original biotite flakes. This assemblage, with a few phenocrysts, composes approximately 75 per cent of the rock. The remaining 25 per cent consists of an interstitial mosaic of cryptocrystalline to microcrystalline quartz and orthoclase, accompanied by much sericite and murky brown limonite.

Many of the dikes show weak gneissic structure. Observed foliation was parallel to the walls and was probably formed by planar flow. However, the possibility of secondary foliation is indicated by minor cataclastic crushing and granulation, seen under the microscope.

The granite porphyry dikes were injected shortly prior to, or simultaneously with, the introduction of the Ithaca Peak porphyry sill and stock, for none were found transgressing these intrusions.

Pegmatite and Aplite

Small dikes, lenses, and irregular masses of pegmatite and small aplite dikes occur within the Ithaca Peak porphyry. They are more abundant in the Mineral Park stock than in the Chloride sill. The rocks are petrographically the same as the simple pegmatites and aplites in the Cerbat complex.

Lamprophyre Dikes

Lamprophyre dikes are located principally within and near the Ithaca Peak porphyry sill and stock, though examples occur as far away

as the eastern base of the Cerbat Range. Lamprophyres are common in the walls of quartz-sulfide veins.

VOGESITE: Vogesite dikes occur in many places, and exposures are excellent immediately northeast of the Tintic mine, just west of the Mary Bell mine, and on the ridge northwest of the Minnesota-Connor mine. Most of the dikes strike northwest. They are usually 5–20 feet thick and normally can be traced for only a few hundred feet. However, the dike just west of the Mary Bell mine is 100 feet thick and extends more than half a mile. The rock is dense and resistant to weathering, and some of the dikes project as irregular ridges, rising 5–10 feet above the surrounding country.

Fresh specimens are dark gray to dark greenish gray, and weathered surfaces are light olive gray to brownish gray. Altered feldspar and hornblende are discernible under a hand lens. The rock almost invariably contains medium- to coarse-grained, rounded crystals of quartz, accompanied in some specimens by crystals of orthoclase feldspar. These minerals appear to be xenocrysts and not original constituents of the rock, though some of the feldspar crystals may be phenocrysts.

Alteration products are normally so abundant that the determination of the rock is difficult. The texture is porphyritic, with a very fine-grained groundmass. Relatively unaltered portions of phenocrysts were determined as both, orthoclase and plagioclase. In the groundmass, brown hornblende needles are abundant. Orthoclase and plagioclase laths, abundant apatite needles, and a few grains of magnetite make up the balance of identifiable primary minerals. There seems to be slightly more orthoclase than plagioclase. Large quantities of pale-green chlorite, with calcite, sericite, kaolin, epidote, and zoisite, have formed as alteration products.

SPESSARTITE: Spessartite dikes are numerous. They range from 18 inches to several feet in width, and most of them strike northwest. One dike, 5–6 feet wide, forms a low ridge immediately north of the Tennessee mine tailings dump. Other examples occur west of the Singlow claim, in the west wall of the Mayflower vein, and in association with the vein immediately south of the Copper Age mine.

The rock is brownish gray on weathered surfaces and dark greenish gray where fresh. Development of large phenocrysts of hornblende and feldspar is characteristic.

A fresh specimen from the dike south of the Copper Age mine was examined under the microscope. Phenocrysts compose 5–10 per cent of this rock. In order of abundance, these consist of brown hornblende, augite, and completely altered feldspar. The hornblende occurs in two generations—crystals completely altered to a mixture of chlorite and epidote, and fresh crystals. The augite is partly chloritized, and the feldspar is thoroughly saussuritized. Fresh feldspar phenocrysts from other specimens were determined as andesine (An_{40}). The relative abundance of hornblende and feldspar phenocrysts varies widely. In some specimens feldspar crystals greatly outnumber the hornblende crystals.

The groundmass consists mostly of laths of andesine, with small amounts of hornblende, augite, magnetite, and apatite.

KERSANTITE: A dike of kersantite occurs in the footwall of a short vein immediately west of the Boulder Dam claim. The dike has a maximum thickness of 5 feet. It is pre-mineral in age.

The fresh rock is grayish brown, and weathered and hydrothermally altered specimens are dark greenish gray. The texture is porphyritic with a microcrystalline groundmass. Most of the phenocrysts are biotite plates, usually well altered. Clear calcite is the chief alteration mineral, and the calcite is without exception partly or totally rimmed by narrow growths of fibrous brown chlorite. Feldspar phenocrysts are altered to calcite that is murky with an aggregate of sericite and other alteration products and typically cut by veinlets of brown chlorite. The nature of the alteration products suggests that the original feldspar was a calcic plagioclase.

The groundmass consists of about 30 per cent fresh biotite, 10 per cent magnetite granules, and the balance chlorite and calcite. A few crystals of sphene are present.

AGE OF LAMPROPHYRES: The lamprophyre dikes were injected after the consolidation of the Ithaca Peak porphyry and prior to the quartz-sulfide vein mineralization. Their time relationships with the volcanic dikes and extrusive rocks are not known. Though the different varieties are thought to be closely related

in origin, and therefore of about the same age, the vogesites seem to have preceded the spessartites, as a dike of the latter cuts a vogesite dike west-northwest of the Singlow claim. The relative position of the kersantite has not been determined.

Tertiary Eruptive Rocks—Summary

Tertiary volcanic strata probably covered the entire Chloride quadrangle. Remnants now occur west of Chloride and discontinuously along the east flank of the range. The oldest eruptive rocks are the Bull Mountain series of flows, tuffs, and breccias predominantly andesitic in composition. These rocks rest upon a surface cut on the Precambrian Cerbat complex. Overlying these basal andesites, with apparent disconformity, are the Kingman series of rhyolite tuffs, breccias, and flows. West of Chloride, on the north side of the mouth of Big Wash, an andesite flow lies unconformably on rhyolite tuff and breccia.

These volcanic extrusions have previously been assigned to the Tertiary by Schrader (1909, p. 34) and Lee (1908, p. 14). There is no proof that all of them are Tertiary, but their inclusion as part of the Tertiary volcanic system, so widespread over the Basin and Range province, is reasonable.

Bull Mountain Series

The Bull Mountain series of flows, tuffs, and breccias occurs along the east side of the range, where only rapid mapping was done, to determine general rock distribution and structure. The series covers several square miles in the southeastern part of the quadrangle. This area is the northernmost extension of the large volcanic mass known as the Kingman Mesa, and the topography is that of a well-dissected, block-faulted table land. The strata reach a thickness of almost 1000 feet in this section, hundreds of feet being exposed in sheer cliff faces. The beds strike predominantly northeast and dip gently southeast. Farther north the Bull Mountain series occurs as isolated small cappings of ridges and buttes, with a maximum thickness of 350 feet.

The series consists mostly of flows of medium-gray to grayish-red purple andesite, with some dark-gray basalt members. Intercalated with the flows, and especially abundant in the basal section, are beds of pink and red scoriaceous breccia and yellow to red tuff and lapilli.

Wherever exposed, the basal beds lie upon an almost flat surface of Precambrian rocks. The maximum relief observed in this surface is not over 50 feet. The surface and the superimposed volcanic strata dip gently away from the mountains. In the southern part of the range the dip is southeast; north of Vock Wash it is northeast. In places the volcanic rocks pass beneath the alluvium of the Hualpai Valley.

Detailed mapping would divide the series into at least two units, for, after an initial period of volcanic activity, the beds in some places were cut by faults and the uplifted portions were removed by erosion. The maximum observed displacement is only 250 feet, and the period of erosion was sufficiently long to restore a flat surface. The remaining beds were then covered by andesitic material. Any tilting which accompanied the faulting must have been slight, as no distinct angular unconformity exists between the lower and upper strata.

In several places the Kingman rhyolite series overlies the Bull Mountain series with concordant attitudes. However, in the section north of Vock Wash, faulting and erosion of the andesitic strata seem to have preceded extrusion of the rhyolite, and the relationships are disconformable, at least in that area. This period of faulting and erosion might represent the intra-Bull Mountain series break, if the upper andesitic beds had not been deposited or had been stripped away prior to the outpouring of rhyolite. More likely however, the faulting and erosion followed the andesitic volcanic activity and represents an intervolcanic break.

Most, if not all, of the volcanic exposures along the east side of the range have been preserved by down-dropping along northwest-trending normal faults, which are younger than the Kingman rhyolite series.

CORRELATION: The Bull Mountain series corresponds to the "older andesite" of Lee (1908, p. 16) and occupies the same stratigraphic position as the Alcyone and Esperanza trachytes, Oatman andesite, and Gold Road latite of the Oatman district in the Black Mountains (Lausen, 1931, p. 27–36).

Kingman Series and Big Wash Andesite

Chloride District

General description.—Small patches of stratiform volcanic rocks occur along the western edge of the Chloride district, on the downdropped side of the Sacramento fault. The beds strike northwest, dip southwest at varying angles, and form a series of isolated low hills along the fault line. The volcanic rocks disappear to the west beneath the alluvium of the Sacremento Valley and tongues of alluvium separate the hills.

Kingman series.—Explosive rhyolitic products of the Kingman series are the oldest exposed volcanic rocks. Several different beds can be distinguished, all conformable, with a strike of N. 50°–60° W. and a dip of 25°–30° SW. Since the Sacramento fault, against which the beds terminate, strikes N. 45° W. and the beds dip southwest, the lowest strata should occur farthest to the north along the fault line. However, this cannot be relied upon, as auxiliary faulting, at high angles to the Sacramento fault, has broken the stratigraphic continuity.

The two northernmost mapped exposures contain a visible section approximately 700 feet thick, composed of coarse reddish breccias. The rocks are distinctly stratified and well consolidated.

In the three exposures to the south several hundred feet of strata must be represented, but the maximum thickness of any one section is about 80 feet. Three rock types are present. The lowest and thickest is a loosely consolidated grayish-yellow to light-brown tuff that is massive in hand specimen. It is light, porous, and can be dented easily with a pick. Essentially a vitric-lithic tuff, it is composed mostly (90 to 95 per cent) of pumice fragments. The remainder of the rock consists of angular, accidental rock fragments. Most of these fragments are gray to reddish porphyritic andesite, but a few are light-gray porphyritic rhyolite. Sparse, isolated crystals of feldspar occur in some specimens.

Overlying the unconsolidated tuff is a bed of welded tuff, 10–15 feet thick. Hand specimens are pale yellowish brown to grayish red, with a streaked and layered structure visible even in tiny fragments. The streaks consist of compressed blebs of black volcanic glass. In some places there are distinct beds of vitrophyre a foot or more thick. The rock is compact and breaks with a rough conchoidal fracture.

Above the welded tuff, but with what seem to be gradational relationships, lies a reddish-brown tuffaceous breccia. A maximum thickness of approximately 30 feet of this breccia is exposed beneath a basalt capping. The rock is composed of fragments of pumice set in a matrix of pumiceous ash and small lapilli. Numerous accidental fragments of andesite are scattered through the rock. The pumice and andesite fragments are essentially the same in composition as those in the welded tuff, and the pumice fragments in the breccia are collapsed into waferlike forms. The chief distinction between the breccia and the welded tuff is degree of coarseness.

Devitrification is more advanced in the pumice of the breccia than in that of the welded tuff, and calcite and tridymite are abundant in the groundmasses of both pumice and andesite fragments.

Big Wash andesite.—A flow of andesite 50–60 feet thick unconformably overlies the pyroclastic rhyolite beds in the northernmost mapped exposures of volcanic rocks. The andesite strikes N. 45° W., essentially parallel to the course of the Sacramento fault, and dips about 15° SW. Platy jointing is well developed in the bed and is expressed through the development of ledgelike outcrops.

History.—All the rhyolite along the Sacramento fault is pyroclastic. Although the exposures are limited, the beds are thick, and extensive volcanism is indicated. Following their deposition, the rhyolite beds were tilted about 15° SW. A long enough period of erosion followed to permit thorough planation. The rhyolite was then unconformably covered by a flow of andesite. Later tilting brought the andesite to an average dip of 15° SW. and the rhyolite beds to about 30°. The late tilting was probably a function of movement on the Sacramento fault. The beds occur on the downdropped side of the fault, and drag could have produced the present attitude of the andesite and part of the dip in the rhyolite. The early tilting of the rhyolite, however, cannot be explained in this fashion, for the beds strike into

the fault. Deformation not related to the Sacramento fault must have occurred, and the initiation of the Sacramento fault must have taken place in a late stage of the regional geologic history.

East Side of Cerbat Range

The remnants of the Kingman rhyolite series which overlie the Bull Mountain series along the eastern flank of the range contain the same types of loosely compacted tuffs, breccias, and welded tuffs that occur in the chloride district. However, the sections are thinner (300 feet is the observed maximum), and rhyolite flows are present in addition to pyroclastics. These flows are light gray, with a vitrophyric texture and a pronounced flow structure. Phenocrysts are sanidine, and a few crystals of orthoclase, oligoclase, quartz, biotite, hornblende, zircon, and sphene. The groundmass is partially devitrified to a cryptocrystalline mosaic.

Deformation has not been so strong on the east side of the range, and the rhyolite beds dip only 6°–8°.

Correlation

The Kingman series corresponds to the young rhyolite series of the Black Mountains, described briefly by Schrader (1909, p. 40) and discussed as the Antelope rhyolite and Sitgreaves tuff by Lausen (1931, p. 37–39). The Big Wash andesite correlates with the "younger andesite" of the Black Mountains (Schrader, 1909, p. 41), and similar strata are mentioned by Lee (1908, p. 16) in the White Hills, Kingman Mesa, and Aquarius mountains.

Quaternary Olivine Basalt

Sheets of olivine basalt 10–25 feet thick cap all the mapped exposures of volcanic rocks along the Sacramento fault. They are flat to gently dipping southwest and lie unconformably on the older rocks. Portions of the basalt layers rest on and interfinger with the Quaternary gravels, and in one place the basalt seems to lie unbroken across the Sacramento fault. Reddish to black scoria is abundant in the basalt areas, and a few volcanic bombs were observed.

The basalt is dark gray where fresh and weathers grayish or dusky brown. It is vesicular and in some places contains amygdules of clear calcite. The texture is porphyritic with phenocrysts of olivine and augite. The groundmass has an intersertal texture and is composed of a dense felt of labradorite laths and scattered grains of magnetite, augite, and olivine, all set in a matrix of dark-brown glass.

The older volcanic rocks had been deposited, block faulting and extensive erosion had determined the present configuration of the mountains, and large amounts of alluvium had collected before the extrusion of basalt. This is shown by the fact that the basalt overlies and interfingers with the alluvium and also overlies the Sacramento fault, which cuts the underlying volcanic rocks. The basalt can be assigned with little question to the Quaternary.

Volcanic Dikes

Rhyolite

Rhyolite dikes are numerous in all the mineralized areas. Small dikes are scattered throughout the Chloride district, many of them in the walls of quartz-sulfide veins. One very large dike in the eastern, highest portion of the area strikes N.25° E., dips steeply west, ranges from 50 to 200 feet wide, and is more than 2 miles long. It crops out in the saddle immediately north of the Lucky Boy mine, runs through the ravine below the Samoa-Brighter Days camp, and passes on south into the Mineral Park district, where it cuts the Ithaca Peak porphyry.

A distinctive zone of thick, steeply dipping rhyolite sheets, striking N. 20° W., cuts through the center of the Ithaca Peak porphyry stock. This zone is as much as a quarter of a mile wide and more than 4 miles long. Thick dikes are exposed at the gateway to the Mineral Park basin and near the Oro Plata mine in Todd Basin. Several prominent rhyolite dikes, striking about N. 40° W. occur near the mouth of Cerbat Canyon. Some of the dikes form strong ridges, whereas a dike that cuts the peak just northeast of Gross ranch has eroded to form a trench through the top of the peak.

The rhyolite is yellowish gray, and most

specimens are megascopically porphyritic. Many dikes are moderately to strongly silicified.

In two specimens examined under the microscope, the rhyolite exhibits a cryptocrystalline to microcrystalline porphyritic texture. The phenocrysts are mostly orthoclase and microcline. Carlsbad twins are common, and some of the phenocrysts are poikilitic with inclusions of quartz, muscovite, apatite, and zircon. Rounded grains of quartz occur as phenocrysts but are not abundant. The groundmass consists of approximately 70 per cent orthoclase and 30 per cent quartz. Scattered crystals of zircon occur in the groundmass, and sericite is abundant in the orthoclase. Veinlets of microcrystalline quartz cut the specimens examined, and the sericite, and perhaps some of the quartz in the groundmass, may be of hydrothermal origin. No mafic minerals were observed in any of the dikes, either microscopically or megascopically.

Chemical analyses would be necessary to determine the composition of the dikes, but the microscopic examination suggests that they are high-potash dikes, similar to those in the Homestake Mine, Lead, South Dakota, described by J. A. Noble (1948).

Rhyolite dikes occur within the Ithaca Peak porphyry and therefore are younger. Their position relative to the lamprophyre dikes was not determined. It is assumed that the rhyolite dikes were intruded in Tertiary time, as part of the Tertiary volcanic system of the Basin and Range province. Correlation with the extrusive Kingman rhyolite series might be possible if compositional similarity could be demonstrated. In view of the high-potash content of the dikes, however, this does not seem likely. The silicification and the truncation and the truncation of some of the dikes by quartz-sulfide veins indicate that the rhyolite is pre-mineral. No rhyolite dikes were observed to cut the veins.

Andesite

A few small andesite dikes cut both the Cerbat complex and the Ithaca Peak porphyry. They range from 6 inches to 4 feet wide, and some of them project as low ridges above the surrounding rocks. None could be traced more than a few hundred feet.

The rock is dark greenish gray to dark gray with a porphyritic texture and a microcrystalline hyalopilitic to intergranular groundmass. Phenocrysts of labradorite (An_{58}), about 5 per cent of the rock, occur as euhedral, slightly resorbed crystals.

About 60 per cent of the groundmass consists of a felt of andesine laths. Interstitial to the laths are green chlorite and grains and rods of magnetite, partly altered to hematite. The shape of some grains suggests that the chlorite was derived from pyroxene.

Some dikes are associated with quartz-sulfide veins and are thoroughly propylitized. Specks of disseminated pyrite, of hydrothermal origin, are common in these altered dikes.

The dikes are possibly related to the extrusive volcanics and are presumed to be Tertiary. Similar to the rhyolite dikes, they are younger than the Ithaca Peak porphyry but of undetermined position with respect to the lamprophyres. No exposures were found in which the sequence of introduction of the rhyolite dikes and andesite dikes could be ascertained. The occurrence of andesite in the walls of veins, with attendant propylitization, indicates a pre-mineral or, at the latest, an inter-mineral age for the dikes.

Basalt

A small olivine basalt dike crops out within the Ithaca Peak porphyry sill, half a mile north of the town of Chloride. A larger basalt dike, 8–10 feet wide, cuts the Cerbat complex in the hillside due west of Merit Spring. The basalt is highly vesicular and amygdaloidal in places. Aside from dating the small dike as post-Ithaca Peak porphyry, no age determinations can be made. The dikes may be Tertiary and related to the basaltic rocks of the extrusive Bull Mountain series, or they may be Quaternary and related to the basalt flows that interfinger with alluvium. The dikes are similar in appearance and composition to flows of both ages.

ALLUVIUM

The alluvium in the region is composed of the detritus that covers the bottoms of the canyons and washes, the fill of the Sacramento

and Hualpai valleys, and the terrace deposits along the walls of many canyons. The detritus now moving along the drainage ways is of course Recent. The terrace deposits represent older alluvium, as does the bulk of the detrital apron flanking the range. The Sacramento Valley apron has been dissected, exposing older debris. The Hualpai Valley apron is only locally dissected, and the older fill is mostly covered with Recent wash.

The terrace beds and dissected portions of the aprons reveal a poorly sorted accumulation of angular to subrounded sand, pebbles, cobbles, and boulders representing all the rocks of the Cerbat Range. Boulders 10–15 feet in diameter are not uncommon, though cobbles 1–5 inches in diameter predominate. A poorly defined stratification exists in some places, as do distinct beds of sand and silt. The deposits are largely unconsolidated, but terraces in the Mineral Park district have been cemented by iron oxides, and white to brown caliche is a weak binding agent in places in the valley fill. In the alluvial veneer west of Mineral Park, mineralizing solutions have bound the debris together in a blanket that forms the Emerald Isle copper deposit. Isolated thin sheets of olivine basalt overlie and interfinger with the alluvium in the Sacramento Valley west of Chloride.

The alluvium began to collect following the initial elevation of the Cerbat Range (probably late Tertiary). Large amounts of detritus accumulated, continually stimulated by progressive lifting of the mountain block, and aggradation in the adjoining basins continued into the Quaternary. Alluvial terraces indicate a late period of aggradation along the canyons within the range. In comparatively recent time a period of degradation was inaugurated, seemingly due to a regional eastward tilt, for rejuvenation effects are well shown in the canyons draining the west slope of the range and in the adjoining Sacramento valley apron, whereas aggradation was only locally interrupted along the east side of the range. Terraces within the mountains are 25–50 feet above present stream levels, and incised washes along the edge of the Sacramento valley are as much as 50 feet below the adjacent surface. The stripping of debris along the western base of the mountains has exposed wide bedrock pediment areas, many of which contain veins. In other areas veins probably remain concealed.

The older alluvial deposits can be correlated with the Temple Bar conglomerate, assigned to the Pleistocene by Lee (1908, p. 63–64).

Geologic Structure

General Features

The Cerbat Range is a massive block composed mostly of crystalline schist, gneiss, and granite. Remnants of folds, however, indicate Precambrian orogenesis.

Foliation is developed to varying degrees in all the plutonic and metamorphic rocks, and three joint systems cut all the rocks.

Faulting that must have occurred in the Precambrian is difficult to detect. Mylonitic zones testify to movement by intense shearing, but the dislocations cannot be dated. Minor movement preceded or accompanied the injection of some of the silicic and mafic dikes, and a period of faulting is evident in the formation of the numerous veins. Four distinct fracture periods occurred during mineralization, and strong post-mineral brecciation is visible in many veins. At least two and possibly four ages of normal faults are shown by remnants of Tertiary volcanic strata. Delineation and elevation of the Cerbat Range has occurred along the youngest of these faults.

Archean Fold System

An orogenic fold system must have once covered this region. Most of the folds have been obliterated by granitization and batholithic intrusion, but a few remnants exist.

The largest fold, an anticline that occupies the western half of the Chloride district, is revealed by the disposition of the crystalline metamorphic and igneous rocks in steeply dipping, concordant, curved layers. This anticline strikes about NE. and plunges about 60° NE. The limbs dip outward at 70°–85°, forming a symmetrical fold. This fold guided the series of igneous intrusions which followed and now form the bulk of the rocks in the structure. During Precambrian time the core of the arch was occupied by coarsely porphyritic Diana

granite, and sheets of pegmatite were injected into the limbs. Later, possibly in the Tertiary, a sill of Ithaca Peak porphyry was introduced into the flanks of the anticline, forming an additional curved igneous layer.

A smaller but well-defined anticlinal remnant is marked by the distribution of septa in the northwest side of the Ithaca Peak porphyry stock in the Mineral Park district. The axis seems to strike north-northeast and plunge 75° NE., but only the nose and northwest limb are preserved, and the axial position is difficult to fix. The fold is at least 2 miles wide, with the strata either vertical or steeply dipping outward. It guided the emplacement of the northwest portion of the Ithaca Peak porphyry stock, and planar-flow structure in the latter developed a mimetic anticlinal pattern. Isolated attitudes suggest that an asymmetrical, almost isoclinal, narrow syncline separates this anticline and the large one in the Chloride district.

In the area including the southwest end of the Ithaca Peak porphyry stock and the northwest portion of the Cerbat district, an anticline is suggested by the configuration of the border of the stock and by the attitudes of foliation, both in the stock and in the rocks of the Cerbat complex. The structure trends northeast and plunges northeast. Minor anticlines and synclines seem to be present, and the entire structure is slightly overturned to the southeast.

A small anticlinal nose, visible only for a few hundred feet, is exposed in the canyon on the north side of Bull Mountain. This fold strikes north-northeast and plunges 12° NE. Another small anticline, observed in the ridge approximately 1½ miles due east of the Stockton mine, strikes north-northeast and plunges 70° NE.

Away from areas where distinct folds are present, the trend of the foliation averages N. 40° E. Aside from vertical layers, about 80 per cent of the observed attitudes dip 75°–80° NW.; the balance dip steeply southeast. This implies the existence of formerly extensive isoclinal folds, the remnants of which are now overturned 5°–15° to the southeast. This overturning may be due in part to original compressive stress. However, at least some and possibly all of it results from eastward tilting of the mountains in Tertiary time. This tilting is also responsible for some of the plunge of the anticline remnants.

No true drag folds were observed, but ptygmatic folds are numerous. These seem to have formed under conditions of active major folding, for their axial planes are approximately parallel to the foliation of the enclosing rocks. As the foliation seems parallel to original bedding, a certain amount of inter-bed shearing must have helped to form the ptygmatic folds.

A consistent feature is the northeast to north-northeast strike of the regional structure lines. Though batholithic intrusion and granitization destroyed much of the continuity, the trend was not disturbed.

Foliation

Cerbat complex.—Foliation in the Cerbat complex ranges from the excellent flow cleavage of biotite schist to faintly gneissic structure in granite. A few areas of massive granite exist, but they are minor. In the anticline remnants, foliation of the schists seems to parallel original bedding, for it wraps around the noses of the folds. The foliate structure possibly developed through a combination of mimetic recrystallization and flow parallel to the bedding, induced by intense hydrostatic and horizontally directed compressive stress and associated extreme temperatures of high-grade regional metamorphism. The foliation of the layers of gneiss, which also parallels the original bedding, could either be due to primary planar flow, be inherited from pre-existing schists by a process of granitization, or be due to compression. Whatever the cause, and more than one may have operated in any one place, the fold pattern was followed faithfully.

On the grounds of this conformity to structure in the remnant anticlines, it is assumed that most of the foliation in the Cerbat complex, especially in the layers of schist, is essentially parallel to original bedding. On this assumption isoclinal folding is deduced in the areas where the foliation has a consistent N. 40° E. strike and a steep dip to the northwest.

Much of the gneiss shows effects of cataclastic deformation, imposed after the consolidation of the rock and after or near the end of orogenic folding. Some of the gneissic foliation is due to this later stress. Transection of earlier

foliation might be expected but was not observed. The continuing or revived compression must have been unchanged in direction and relieved so that younger foliation merely added to the intensity of older or was formed parallel to it. In plutons, such as the Diana granite, secondary foliation in the interiors of the masses has, in most places, a northeast strike and steep dip, parallel to the regional grain.

Secondary lineation is present in the hornblende schists. In the few observations made, the long axes of the hornblende crystals seemed to parallel the axes of associated folds. No attempt was made to apply methods of structural petrology to the schists. Such a study would undoubtedly add much information on the tectonic history and on the nature of the deforming forces in Precambrian time.

Ithaca Peak porphyry.—Primary planar flow structure is well developed in many places along the borders of the Ithaca Peak porphyry sill and stock. The flow structure parallels the contacts and is characteristically strongest immediately adjacent to the contacts. This parallelism is its diagnostic feature. Primary foliation is especially well shown between the anticlinal septa in the northwest portion of the Mineral Park stock. Border areas, however, may be massive or only slightly gneissic.

Compressive forces, possibly contemporaneous with and definitely outlasting consolidation of the Ithaca Peak porphyry, caused secondary foliation which differs in degree from place to place. It is best developed in the Chloride district sill. Local secondary foliation occurs in the Mineral Park stock, but the bulk of the intrusion is massive. The compressive forces were from the northwest or southeast, for most of the foliation strikes about N. 50° E., and is vertical or has steep dips. This predominating trend is shown in Figure 2, a point diagram prepared in the manner outlined by Billings (1942). This technique, which utilizes the plotting methods of structural petrology, shows the poles of perpendiculars to surfaces such as folia, joints, and veins. Perpendicular surfaces are represented by points on the circumference of the plotting circle, and horizontal surfaces are represented by a point in the center. Both primary and secondary foliation have been plotted. This has not altered a dominant northeast strike, probably because the contacts of the long western limb of the Chloride sill, as well as the contacts of the northwest part of the Mineral Park stock, have a northeast strike, and the primary foliation parallel to these contacts provided a large portion of the total observations.

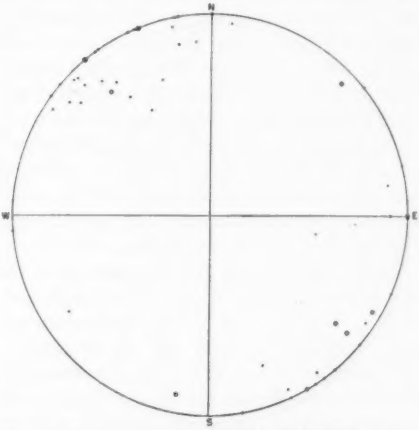

FIGURE 2. ATTITUDES OF ITHACA PEAK PORPHYRY FOLIATION SURFACES

Point diagram showing poles of perpendiculars to 53 foliation surfaces in the Ithaca Peak porphyry. Plotted on upper hemisphere. Equal area projection. Dots represent surfaces in the Chloride district sill; small circles represent surfaces in the Mineral Park stock.

Joints

Cerbat complex.—Several types of stress created the fracture systems in the Precambrian. Horizontal compression is evidenced by the folding; strong hydrostatic pressure would be a factor in the high-grade regional metamorphism; and thermal expansion and contraction necessarily would be associated with the intrusion of the igneous rocks. Stresses of all later geologic periods also might develop fractures in these ancient rocks.

The resulting joints are therefore complex but can be resolved into three trends, exhibited in both the early schists and in the later granite and gneiss. This leads to the conclusion that directional stress, having a uniform orientation both in the Precambrian and later, was the principal factor in the establishment of the joint pattern. In field observations the joints

were recorded as strong, moderate, or weak—"strong" if they seemed to extend beyond the area of observation and there were numerous parallel joints, "weak" if they seemed to be

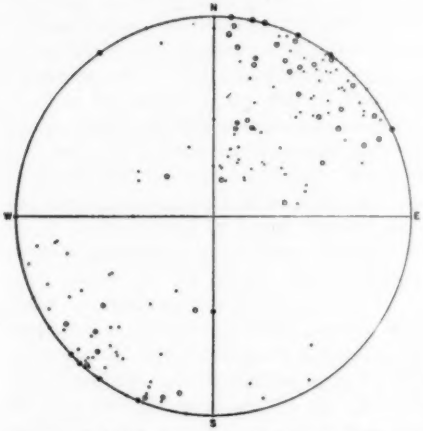

FIGURE 3. ATTITUDES OF "STRONG" JOINTS

Point diagram showing poles of perpendiculars to 175 "strong" joints. Plotted on upper hemisphere. Equal area projection. Dots represent joints in Cerbat complex; small circles represent joints in Ithaca Peak porphyry.

small and there were few parallel joints, "moderate" if between these extremes. In most places, moderate joints were recorded only where stronger joints were present for comparison. The terms are relative, and the classification of a joint can change with distance. However, correlation was obtained between directional trends and relative strength. The strongest joints strike about northwest and in general dip very steeply. This is demonstrated by the massing of points in the northeast and southwest quadrants of Figure 3. Joints with dips to the northeast are most numerous, but a set that dips to the southwest is noticeable. The moderate joints (Fig. 4) display less consistency in direction, but a northeast strike is favored. Steep southeast dips are most numerous, but a set that dips to the northwest is present. The weak joints are scarce and give a scattered point diagram (Fig. 5), but where three different joints occur together, field observations show a tendency for the weakest set to strike north. Dips are either east or west.

The orientation of folding and foliation in-

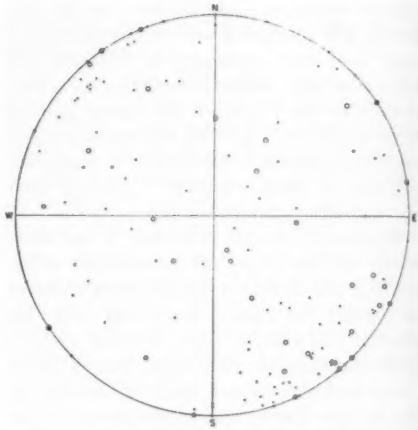

FIGURE 4. ATTITUDES OF "MODERATE" JOINTS

Point diagram showing poles of perpendiculars to 136 "moderate" joints. Plotted on upper hemisphere. Equal area projection. Dots represent joints in Cerbat complex; small circles represent joints in Ithaca Peak porphyry.

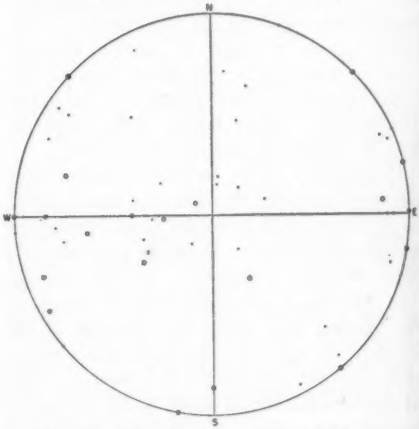

FIGURE 5. ATTITUDES OF "WEAK" JOINTS

Point diagram showing poles of perpendiculars to 54 "weak" joints. Plotted on upper hemisphere. Equal area projection. Dots represent joints in Cerbat complex; small circles represent joints in Ithaca Peak porphyry.

dicates that the compressive forces affecting the Precambrian rocks were oriented about northwest-southeast. Under such a stress, the theoretical fracture planes that should develop in the rocks agree well with many of the observed joints. Tension fractures should form parallel to the direction of compression, as represented

by the northwest-trending, strongest set of joints. Shear fractures should be directed northeast and north, as are many of the moderate and weak joints. The wide variations in attitude in the moderate and weak joints emphasize the actual complexity of stress and strain conditions, but the uniformity of the strong joints suggests that they did originate as tension fractures due to compressional deformation.

The joint pattern was of major importance in later geological events. The fractures served as avenues of injection for numerous silicic and mafic dikes and were a prime factor in determining the location and orientation of the quartz-sulfide veins.

Ithaca Peak porphyry.—Joints in the Ithaca Peak porphyry sill and stock follow the same trends as those that cut the Cerbat complex (Figs. 3, 4, 5). Some of the joints must be genetically related to the intrusion and solidification of the magma. However, northwest-southeast compression, similar to the Precambrian forces, established secondary foliation and is believed responsible for most of the joints.

Faults

General statement.—The oldest recognizable faulting has left zones of mylonite in the Cerbat complex. The mylonite zones have been assigned a Precambrian age, which may or may not be valid. They represent localized intense shearing parallel to the northeast structural trend of the Cerbat complex.

Beginning with the intrusion of the Ithaca Peak porphyry, several ages of faulting are evident. Minor dislocations accompanied the periods of injection of granite porphyry dikes, lamprophyres, and rhyolite and andesite dikes. The maximum displacement for any of these was observed in the ridge northeast of the North Georgia mine. Here a vogesite dike offsets a diabase dike 100 feet horizontally. Dike injection and associated faulting followed the already established jointing trends in the Cerbat complex and the Ithaca Peak porphyry.

Mineralized faults.—The most obvious faults are those along which veins have formed. Detailed mapping in the Chloride district has revealed a large number of veins (Thomas, 1949, Fig. 2). Most occur in the eastern half of the district, where the veins in general strike northwest. The vein system is continuous to the south through the Mineral Park, Stockton Hill, and Cerbat districts and is fundamentally unchanged in pattern, as more than 80 per cent of the observed veins strike west-northwest or northwest. In the central part of the Chloride district the veins trend north, parallel with the intrusive contacts of the eastern limb of the Ithaca Peak porphyry sill. In the western part the veins are less numerous and have no uniform trend. An east-west orientation is characteristic in the center of the Diana granite. Some of the veins intersect, and many are joined along the strike. Vertical to steep dips characterize most of the veins. Northeast dips are most numerous, but there are many southwest dips. In the steeper veins reversals of dip are not uncommon.

Striations in the vein walls indicate that displacement has almost invariably been oblique. The prevalent slickensided surfaces, large amounts of gouge, and rolled and crushed fragments of wall rock show that shearing stress must have been important. Most are believed to be normal faults, though some seem definitely reverse. In steeply dipping oblique-slip faults the distinction between normal and reverse movement has little significance, as a change in dip of a few degrees can easily result in a change in classification.

It is impossible to determine the amount of movement in most places. A few mineralized faults offset dikes of pegmatite and lamprophyre, however, and a horizontal displacement of 120 feet was measured along a vein that offsets a lamprophyre in the northwest corner of Sec. 33, T. 23 N., R. 18 W. Vertical movement could not be determined.

Five periods of activity are reflected in the mineralized faults. The first opened the fissures and gave access to mineralizing solutions. Three intramineral fracture periods followed, then strong post-mineral fracturing.

The mineralized faults postdate Tertiary (?) rhyolite and andesite dikes.

Transverse faults.—Steeply dipping, intramineral cross faults cut some of the veins and produced small offsets. In the Tennessee-Schuylkill mine, a fault striking N. 50° E. and dipping 55°–60° N. has offset the main vein a maximum of 30 feet horizontally.

A large post-mineral transverse fault cuts

several veins in the eastern part of the Chloride district. It strikes N. 55° E., is essentially vertical, and can be traced for three-quarters of a mile. The north end of the Payroll vein is terminated against this fault. The North Georgia vein, on the opposite side of the fault, is similar in size and appearance and seems to be a continuation of the Payroll vein. This indicates a left-lateral offset of about 500 feet. The fault surface is exposed on the north side of the wash, a quarter of a mile southwest of the Payroll mine shaft. In the fault exposure a 1-foot width of unmineralized gouge and breccia occurs.

Normal faults.—Evidence of normal faulting is common where remnants of volcanic strata occur (Fig. 6). Four periods of faulting have been inferred from the relations of volcanic exposures. The earliest took place within the period of extrusion of the Bull Mountain andesitic series. The rhyolite which lies disconformably upon the Bull Mountain series suggests a post-Bull Mountain series stage, and the tilting of rhyolitic strata along the west side of the range reflects a post-Kingman series stage. A late stage is shown by the breaking and tilting of the Big Wash andesite. Only relative ages can be assigned to the faulting, and the two middle periods may be questioned. Possibly these periods correspond to the four periods of basin-range faulting established by Sharp (1939, p. 902) in the Ruby-East Humboldt range of northeastern Nevada and by Ferguson and Cathcart (1924) in the Hawthorne and Tonopah quadrangles of western Nevada.

The volcanic rocks in the Chloride quadrangle are most abundant along the east side of the range. Therefore most of the observed faults occur on that side, since the faults are visible only within the volcanics. They undoubtedly continue within the basement rocks, and similar faults probably exist outside the volcanic areas. Most of the faults strike north-northwest in echelon. The others trend northeast in echelon. The faults are steeply dipping, and dislocation seems to have been dip-slip. Total displacement could not be determined, but in many of the faults it probably was not more than a few hundred feet. The fault that passes just west of the volcanic-capped mesas north of Vock Wash, however, has displaced the volcanic cap rock a minimum of 700 feet vertically. The alignment of unusually straight canyons and ridge slopes, north and south of Vock Wash, suggests a length of 10 miles for this fault. It passes just west of the Neal ranch buildings and is here termed the Neal ranch fault. It possibly continues and bounds the west side of the lava-capped mesa which is north of Neal ranch and outside the border of the quadrangle. Most of the faults along the east side of the range, including the Neal ranch fault, are believed to be post-volcanic.

The Sacramento fault, west of the Chloride district, is also revealed by the displacement of a series of volcanic strata. It strikes N. 48° W. and dips approximately 60° S. From the point where it disappears beneath alluvium southwest of the district, it was mapped to the northwest for 2¼ miles. The fault continues an unknown distance beyond the area mapped. In Figure 6 the line of faulting is continued to the south by a dotted line. The position of this line was located just west of the pediment area along the front of the range. Faulting cannot be proved along this line, but the continuation of the Sacramento fault, or the existence of similar normal faults, is suggested by the truncation at the mountain front of the internal structure of the range. The lack of outlying ridges and buttes further suggests that the faulting occurred along a single line or narrow zone, in contrast to the echelon pattern along the east side of the range.

The amount of displacement along the Sacramento fault cannot be determined accurately. A calculation of possible movement was made by using a 300-foot horizontal offset of the contact between the Ithaca Peak porphyry and the Cerbat complex. By neglecting any effects of auxiliary-faulting and assuming only dip-slip motion, a net vertical displacement of 3165 feet and a net horizontal separation of 1830 feet were obtained. That this may be approximately correct is indicated by the average relief of 3000 feet in the northern part of the Cerbat Range.

Though the total displacement along the Sacramento fault must have taken place by intermittent movement, it seems to be one of the youngest geologic features in the region, for it was not in existence during the outpouring of Tertiary lavas. On the basis of the assumed

Tertiary age of the rhyolites and andesites, the faulting probably started in late Tertiary and

faulting in the Great Basin seems to have been latest Tertiary and Quaternary". Evidence of

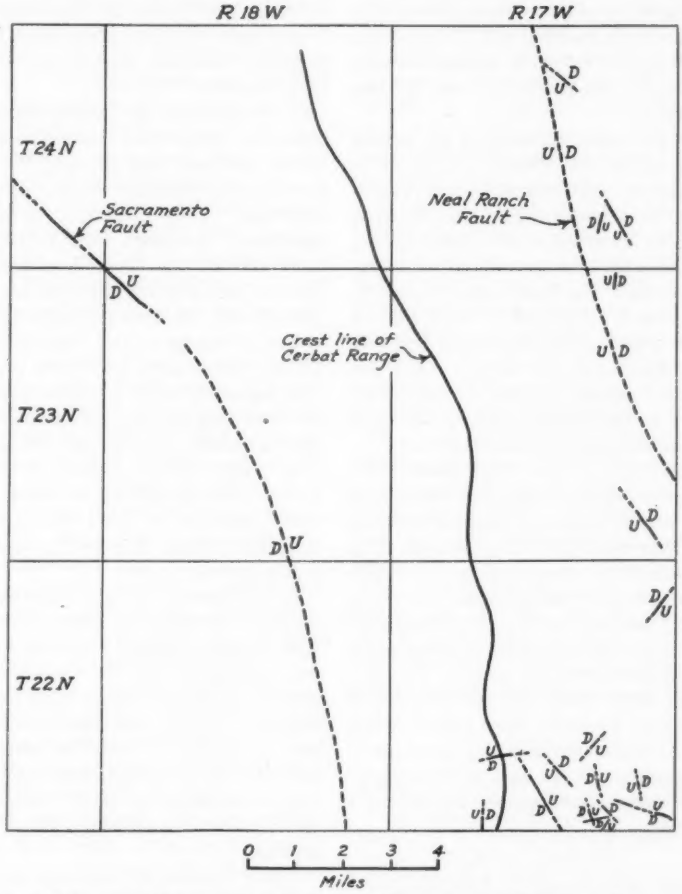

FIGURE 6. TREND AND DISTRIBUTION OF NORMAL FAULTS IN THE CHLORIDE QUADRANGLE

Inferred faults shown by dashed lines

possibly extended into Quaternary time. Even late Quaternary activity is suggested by the eastward tilting and rejuvenation of the mountain block, which has caused the development of Recent gravel terraces. The most logical cause of such tilting would be vertical displacement along the Sacramento or other faults on the west side of the range. R. P. Sharp (1939, p. 905), summarizing studies of basin-range structure, states that "the period of maximum block

block faulting in the Cerbat Range is in accord with this generalization.

Auxiliary faults.—Two auxiliary faults, presumably vertical, cut the hanging-wall block of the Sacramento fault and join the Sacramento fault at angles of about 65°. They bound a segment of the Kingman rhyolite series and indicate breaking and differential movement of the hanging-wall block. The segment of rhyolitic strata has been dropped down relative to

either side and is bounded on three sides by Ithaca Peak porphyry.

Relative downdropping between vertical auxiliary faults has also preserved narrow strips of basalt along the footwall side of the normal fault southeast of Neal ranch. No measurement of displacement was possible on any of these auxiliary faults.

Emerald Isle fault.—Evidence of late normal faulting exists at the Emerald Isle mine in the Sacramento valley west of Mineral Park. Here both the bedrock pediment and the overlying blanket of alluvium have been faulted, and the fissure has been mineralized by copper-bearing solutions. The fault strikes N. 30° E. and dips between vertical and 45° N. The fissure vein in the gravels is as much as 12 feet wide and has distinct walls. The amount of displacement was not determined, but the occurrence of bedrock in the footwall, next to gravel in the hanging wall, indicates a normal fault.

Fault-block form of the Cerbat Range.—The delineation and uplift of the present mountains has been a consequence of large normal-fault activity. Maximum dislocation occurred along the west side of the range, resulting in an eastward-tilted block. In the northern half of the quadrangle, however, between the Sacramento and Neal ranch faults, the range is an eastward-tilted horst. The structural grain of the basement rocks, which trends northeast, is truncated almost at right angles by these normal faults. Differential vertical stresses must have been of major importance, in contrast to the compressive forces that characterized the earlier history of the region.

Geologic History after Precambrian Time

Following the Precambrian, a large gap occurs in the geologic record. Several thousand feet of Cambrian, Mississippian, and Pennsylvanian sediments overlie the basement crystalline complex in the Grand Wash Cliffs, only 12–15 miles east of the Chloride quadrangle. In the Cerbat Range similar strata are lacking. With large thicknesses of Paleozoic rocks so close at hand, their former extension over the Cerbat complex would be expected. They might have been much thinner, though, as the Chloride quadrangle is on the northeast edge of Schuchert's (1910) "Ensenada", a Paleozoic land area or positive element. If Paleozoic sediments were present, removal has been thorough. Schrader (1909, p. 30) mentions limestone from a well northwest of Kingman which he thought might be Paleozoic. Nothing similar has since been recorded.

No features can be related definitely to the Mesozoic. If the region received Mesozoic sediments they have been removed. The Laramide revolution is reflected in the Muddy and Virgin mountains, in the region north of the Chloride quadrangle (Longwell, 1928, p. 126). It may have affected the Chloride quadrangle, and features such as secondary foliation and mylonitic zones in the Cerbat complex might be due to late Cretaceous or early Tertiary compressive forces. Furthermore, the Diana granite could be of Laramide age. It is felt that these features fit better into the Precambrian sequence, and they have been so classed, but the classification is arbitrary. There is nothing, aside from alluvium and its associated sheets of olivine basalt, that can be dated with any degree of assurance. Specific ages can be assigned only with a question mark.

In the Tertiary (?), the Ithaca Peak granite porphyry intruded the Chloride and Mineral Park districts, guided by pre-existing structure and accompanied by the profuse injection of granite porphyry dikes, especially in the Chloride district. Northwest-southeast compressive forces, in operation possibly during and definitely after emplacement and consolidation of the granite porphyry, produced different degrees of secondary foliation in the intrusions.

Minor amounts of pegmatite and aplite are associated with the Ithaca Peak porphyry, and lamprophyre dikes are numerous. It is common for all the dikes, including the granite porphyry dikes, to occur in composite bodies. Many old fracture lines, previously invaded by pegmatite and diabase, were successively reopened to granite porphyry and lamprophyres, and in some places to rhyolite and to vein-forming solutions.

This entire region probably once was covered by a great thickness of pyroclastic volcanic debris and lava flows of Tertiary (?) age. Andesitic flows, tuffs, and breccias of the Bull Mountain series represent the first stage of

volcanism. A period of normal faulting and erosion occurred between the eruption of the lower and upper beds, and erosion and probable normal faulting followed the accumulation of the series. Succeeding volcanism came from silicic magma, and the Kingman series of rhyolite flows, tuffs, and breccias was deposited disconformably upon the lower andesites. Tilting of the rhyolites, probably due to faulting, and truncation by erosion took place on the west side of the range but is not indicated on the east side. A later eruption unconformably capped the Kingman rhyolite with the Big Wash andesite, to end the general period of volcanism. Numerous rhyolite and a few andesite dikes of Tertiary (?) age cut the basement rocks.

Sulfide mineralization postdated the injection of the rhyolite and andesite dikes. The first period of mineralization was localized almost entirely within the Mineral Park stock of Ithaca Peak porphyry. A large part of the stock was broken into closely spaced fractures which were utilized by hydrothermal mineral solutions. The result was a low-grade, disseminated sulfide deposit.

After an unknown interval a second and more extensive period of mineralization occurred. The solutions followed fissures, mostly localized along northwest-trending old fractures. The result was a system of quartz veins that contain sulfide ore minerals. Veins of this period cut across the disseminated sulfide deposit and are the only ore deposits of the Chloride, Cerbat, and Stockton Hill districts.

After the conclusion of volcanic activity, presumably in the late Tertiary, the present mountains began to form. The Sacramento normal fault, or fault system, was formed, and the east side of the fault was progressively elevated relative to the west side. A total vertical displacement of several thousand feet was eventually reached. The rising mountain block was also tilted so that it sloped 5°–10° E. Contemporaneous with activity along the Sacramento fault, normal faults in echelon, most of them striking northwest, disrupted the rocks along the east side of the block. Vertical displacements were not as large as on the Sacramento fault, but the Neal ranch fault had significant dip-slip movement and established a tilted horst structure for part of the mountains.

Large quantities of erosional debris were shed from the rising block, and a broad pediment was cut along the base of the range. At a late stage, alluvium had begun to choke the canyons draining the mountains. The western side of the block was then elevated once again, and the streams are now downcutting, having left terraces up to 50 feet above their present grades. This rejuvenation is believed to stem from late Quaternary movement along the Sacramento or related faults.

In late Quaternary time, after the bulk of the alluvial fill had accumulated in the Sacramento valley, and after cutting of the pediment, another period of mineralization occurred. Mineralizing solutions ascended along a northeast-striking normal fault fissure, which extended out of bedrock into alluvium at the site of the present Emerald Isle mine. Chrysocolla and other copper silicate minerals were deposited both in the fault fissure and in the surrounding alluvium. This mineralization must be later than the main period of active movement along the faults bounding the range. It, may be earlier than the last rejuvenation however.

The last geological event, apparently of Recent age, was the extrusion of sheets of olivine basalt. These overlie and interfinger with alluvium, and in one place basalt seems to lie across the Sacramento fault.

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SILICEOUS SEDIMENTS OF OUACHITA FACIES IN OKLAHOMA

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ABSTRACT

Certain sedimentary rocks cropping out in the Ouachita Mountains of southeastern Oklahoma and western Arkansas are unusual in that they contain large amounts of silica disseminated through a relatively thick series of beds which are laterally continuous over large areas. They range in age from Ordovician to Pennsylvanian and may be divided into (1) sedimentary rocks of the main anticlinorium of the Ouachita Mountains in southeastern Oklahoma and southwestern Arkansas, and (2) sedimentary rocks exposed in the fault blocks lying between the central Ouachita Mountains and the Arbuckle Mountains and McAlester Basin. The former group consists largely of chert and siliceous shale of typical geosynclinal facies; the latter group consists largely of chert and siliceous limestone intermediate in lithology and fabric between rocks of geosynclinal facies and those of foreland facies exposed near by in the Arbuckle Mountains.

The siliceous sedimentary rocks of typical geosynclinal facies comprise the Bigfork chert (Ordovician); the Arkansas novaculite (Devonian-Mississippian); the siliceous shales of the Stanley (Mississippian), Jackfork (Pennsylvanian), and Atoka (Pennsylvanian) formations; and the Hatton tuff lentil (Mississippian). Siliceous sedimentary rocks of atypical geosynclinal facies exposed in the fault blocks are the Pinetop chert (Devonian), the Woodford chert (Devonian ?), and the Chickachoc chert (Pennsylvanian).

Although there are conspicuous lithologic differences between these siliceous sedimentary rocks of Ouachita facies, it is considered that most of the silica was supplied by extrusive vulcanism and submarine weathering of volcanic ash. Furthermore, all the siliceous sediments apparently were deposited when relatively little clastic material was being supplied to the "Ouachita geosyncline". Subsequent to their deposition, the siliceous sediments have been modified by epigenetic redistribution of part of their silica, resulting partly from metamorphism accompanying the Ouachita orogeny and partly from surface exposure.

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INTRODUCTION

General Statement

Certain sedimentary rocks of the Ouachita Mountains in southeastern Oklahoma and western Arkansas are very unusual in that they contain a large amount of silica disseminated through a thick and relatively continuous series of beds. Such a thick section of highly siliceous sedimentary rocks ranging in age from Ordovician to Pennsylvanian and extending over a large area are of scientific importance as well as a major element in the geology of the Mid-Continent region. Siliceous sedimentary rocks of Ouachita facies consist of unusual rock types such as radiolarian chert, spiculitic chert, sapropelic chert, novaculite, and siliceous shale. A number of detailed geologic investigations and a considerable volume of geologic literature has been devoted to them, yet the question of their origin is not yet settled. Except for Honess' (1923) investigations in McCurtain County, Oklahoma, petrographic studies of the sedimentary rocks of the Ouachita Mountains of Oklahoma have been scanty and superficial. The most detailed geologic mapping in the western part of the Ouachita Mountains of Oklahoma is that of Hendricks *et al.* (1947), yet the petrography of the sedimentary rocks of this area has not been described.

The primary purpose of this paper is to present an integrated petrographic study of the siliceous sediments of the Ouachita Mountains of Oklahoma. Sedimentary rocks of pre-Stanley (Mississippian) age outcrop at three places in the Ouachita Mountains of Oklahoma (Fig. 1): (1) Black Knob Ridge, Atoka County, Oklahoma, (2) The Potato Hills, Latimer and Pushmataha counties, Oklahoma, and (3) McCurtain County, Oklahoma. Petrographic descriptions of the siliceous sedimentary rocks from Black Knob Ridge are emphasized here; the other two areas are discussed in less detail.

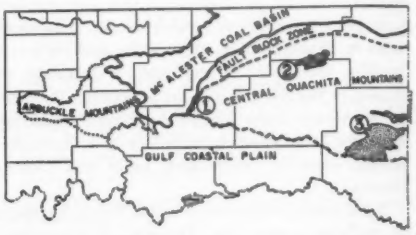

FIGURE 1.—OUTLINE MAP OF SOUTHEASTERN OKLAHOMA SHOWING LOCATION OF AREAS DISCUSSED IN TEXT

(1) Black Knob Ridge; (2) Potato Hills; (3) McCurtain County. Stippled areas are those in which pre-Stanley (Mississippian) sediments are exposed. Adapted from Hendricks, Knechtel, and Bridge (1937).

The sediments discussed herein (Table 1) include the Bigfork chert (Ordovician), the Pinetop chert (Devonian), the Woodford chert (restricted) (Late Devonian or Early Mississippian), the Arkansas novaculite (Devonian and Mississippian), the Hatton tuff lentil of the Stanley shale (Mississippian), and the siliceous shales of the Stanley (Mississippian), Jackfork (Pennsylvanian), and Atoka (Pennsylvanian) formations. Siliceous shales in the Missouri Mountain shale (Silurian) are identical to the siliceous shales of the upper part of the Arkansas novaculite at Black Knob Ridge and are not discussed separately.

Location of Samples

Most of the samples studied were obtained from the vicinity of Black Knob Ridge, Oklahoma. Less extensive collections were made in the Potato Hills, Oklahoma, and in McCurtain County, Oklahoma. A few samples of the Arkansas novaculite and Hatton tuff lentil were collected in western Arkansas.

TABLE 1.—SEQUENCE OF STRATA CHARACTERISTIC OF EACH OF THE STRUCTURAL BLOCKS OF THE WESTERN PART OF THE OUACHITA MOUNTAINS IN OKLAHOMA.*

Age		McAlester Coal Basin	Block Southeast of the Choctaw Fault	Block Southeast of the Katy Club Fault	Block Southeast of the Pine Mountain Fault	Block Southeast of the Ti Valley Fault
Pennsylvanian		McAlester shale Hartshorne sandstone Atoka formation Wapanucka limestone	Atoka formation Wapanucka limestone	Atoka formation Chickachoc chert	Atoka formation	Atoka formation Johns Valley shale
Pennsylvanian		Springer formation	Springer formation	Springer formation	Springer formation	Jackfork sandstone
Mississippian		Caney shale Sycamore limestone	Caney shale	Caney shale	Caney shale Sycamore limestone (?)	Stanley shale
Devonian (?)		Woodford chert			Woodford chert	Arkansas Novaculite
Devonian	Hunton Group	Bois d'Arc limestone Haragan shale			Pinetop chert Unnamed limestone	
Silurian	Hunton Group	Henryhouse shale Chimney Hill limestone				Missouri Mountain, shale
Ordovician		Sylvan shale Fernvale limestone Viola limestone Simpson group —Arbuckle group				Polk Creek shale Bigfork chert Womble shale
Cambrian		Reagan sandstone				

* Adapted with slight modification from that of Hendricks, Gardner, and Knechtel (1947).

PERIOD	FORMATION	COLUMNAR SECTION	THICKNESS (FEET)	CHARACTER OF ROCKS
	Major Unconformity			
PENNSYLVANIAN	ATOKA		1500-9500	SANDSTONE, SANDY SHALE AND SHALE OF VARIABLE CHARACTER
PENNSYLVANIAN	JACKFORK		3500-6600	SANDSTONE, MASSIVE, MEDIUM TO FINE-GRAINED INTERBEDDED WITH GRAY TO GREEN SHALE
MISSISSIPPIAN	STANLEY		3500-6000+	SHALE, DARK-COLORED, INTERBEDDED WITH GRAY FINE-GRAINED ARGILLACEOUS SANDSTONE. SEVERAL BEDS OF VOLCANIC ASH NEAR BASE
	HATTON TUFF LENTIL		(0-90)	
	HOT SPRINGS SANDSTONE		85-200	SANDSTONE, HARD, QUARTZITIC, CONGLOMERATIC AT BASE. CROPS OUT ONLY IN ARKANSAS.
	Unconformity			
DEVONIAN-MISS.	ARKANSAS NOVACULITE		250-900	UPPER MEMBER - CALCAREOUS, WHITE. MANGANIFEROUS NOVACULITE: MIDDLE MEMBER - BLACK WITH INTERBEDDED BLACK SHALE LOWER MEMBER - WHITE, MASSIVE
	Unconformity			
SILURIAN	MISSOURI MT.		0-300	SHALE AND SLATE, RED, GREEN AND BLACK, FISSILE
	Unconformity			
	BLAYLOCK		0-1500	SANDSTONE, GRAY TO GREEN AND DARK FISSILE SHALE VEINS OF SMOKY QUARTZ AT PLACES.
	Unconformity			
ORDOVICIAN	POLK CREEK		100-200	SHALE, BLACK, GRAPHITIC, CONTAINS GRAPTOLITES
	BIGFORK		500-700	CHERT, HARD, GRAY TO BLACK, BLACK SHALE AND DARK-COLORED, FOSSILIFEROUS, CHERTY LIMESTONE
	WOMBLE		250-1000	SHALE, BLACK TO GREEN, WITH THIN LAYERS OF LIMESTONE AND ARGILLACEOUS SANDSTONE IN ARKANSAS. PREDOMINANTLY SCHISTOSE, MICACEOUS, CHLORITIC SANDSTONE IN Mc CURTAIN CO., OKLA.
	BLAKELY		0-500	SANDSTONE, GRAY, QUARTZITIC SOME CALCAREOUS SANDSTONE AND MINOR AMOUNTS OF BLACK AND GREEN CLAY SHALE SMOKY QUARTZ VEINS AT PLACES.
	Unconformity			
	MAZARN		1000±	SHALE, BLACK AND GREEN, BANDED. THIN LAYERS OF GRAY SANDSTONE AND IN SOME AREAS, THIN LENSES OF BLUISH-BLACK LIMESTONE
CAMBRIAN	CRYSTAL MOUNTAIN		500-850	SANDSTONE, MASSIVE, LIGHT-GRAY, CALCAREOUS TO QUARTZITIC. CHERT CONGLOMERATE AT BASE IN Mc CURTAIN COUNTY. IN SOME AREAS CONTAINS MANY QUARTZ VEINS AND CRYSTALS.
	Unconformity			
	COLLIER		200-500	SHALE, BLACK, GRAPHITIC AND DARK-COLORED SILICEOUS LIMESTONE. SOME DENSE BLACK CHERT IN ARKANSAS.

FIGURE 2.—GENERALIZED COLUMNAR SECTION OF SEDIMENTS OF OUACHITA FACIES

Terminology used is based on field lithology. Both thicknesses and lithology are typical of McCurtain County, Oklahoma and western Arkansas. Adapted from columnar sections of Honess (1923) and Shreveport Geological Society 1948 Guidebook.

General Structural Setting

The Ouachita Mountains of southeastern Oklahoma and western Arkansas have an area of about 12,000 square miles, of which slightly more than half is in Arkansas. The area is one of complex structure and stratigraphy. At least four major thrust faults are present in the Ouachita Mountains of Oklahoma. Each major fault block has been broken by numerous small high-angle reverse faults and by less numerous tear faults and normal faults. A sequence of sedimentary rocks ranging from Cambrian to Pennsylvanian is exposed (Fig. 2); the thicknesses and lithology are typical of the central anticlinorium of the

Ouachita Mountains in western Arkansas and southeastern Oklahoma.

Discussion of Term "Ouachita Facies"

The term "Ouachita facies" is used widely by geologists in describing certain Paleozoic sediments of the Midcontinent Region. To many geologists the term is synonymous with "geosynclinal" facies, as opposed to "foreland" or "Arbuckle facies". However, most of the Early Paleozoic formations of the Ouachita Mountains were deposited between the foreland and the axis of active geosynclinal downwarping, are not unusually thick, and lack some of the characteristics usually associated with sediments of "geosynclinal" facies. Consequently, it is here proposed to define sediments of Ouachita facies as those lithologically similar and stratigraphically equivalent to sediments exposed in the Ouachita Mountains of Oklahoma and Arkansas.

Method of Presentation

This paper is divided into three main sections concerning: (1) the stratigraphy and petrography of the siliceous sediments of the central Ouachita Mountains; (2) the stratigraphy and petrography of the siliceous sediments of the faulted zone between the McAlester Basin and the central Ouachita Mountains; (3) the origin of the major units of siliceous sedimentary rocks of the central Ouachita Mountains.

Most of the information on stratigraphy was obtained from published articles by Miser (1917), Miser and Purdue (1929), Honess (1923), Harlton (1938), and Hendricks and various collaborators (1937; 1947).

Acknowledgments

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Bigfork Chert (Ordovician)

General Statement

The Bigfork chert of Ordovician age, the oldest siliceous sedimentary rock of the Ouachita Mountains, crops out at Black Knob Ridge, in the Potato Hills, and in McCurtain County, Oklahoma. In general the Potato Hills section is similar to that at Black Knob Ridge and will not be discussed separately.

Black Knob Ridge

Stratigraphy.—A brief but comprehensive summary of the stratigraphy of the Bigfork chert at Black Knob Ridge has been published by Hendricks *et al.* (1947, sheet 1). They state:

> "The Bigfork chert overlies the Womble shale with an apparently conformable contact, and in some exposures the contact appears to be gradational. The formation is consistently about 600 feet thick, and may be divided into two parts on the basis of lithology. The parts are not of member rank, because as one thickens the other thins; the entire formation maintains a uniform thickness, thus indicating that the line of division between the two occurs at different stratigraphic positions in different sections. The lower part is 410 to 510 feet thick and is characterized by light-gray to light-brown chert in beds mostly 2 to 4 inches thick, although some are as much as a foot thick. Gray limestone forms beds, stringers, and lenses in the chert and locally constitutes about one-fifth of the lower part of the formation. The limestone for the most part is finely crystalline, but a few thin beds in each well-exposed section are coarsely crystalline. The upper part of the formation, 100 to 190 feet thick, consists of chert in beds ranging from 1 inch to 2 feet thick in which beds of black paper shale as much as 2 feet thick are intercalated. The chert in this zone is finely laminated parallel with the bedding and is very dark brown to drab.
>
> The fauna of the Bigfork is composed of graptolites, small brachiopods, trilobites, conodonts, spicules of various sorts, and small bodies which may be radiolarians. The graptolites occur at several horizons throughout the formation, and for the most part they are conspecific with those found in the Viola limestone of the Arbuckle region. They are of Trenton age . . ."

Petrography.—Three types of siliceous sedimentary rocks predominate in the lower part of the Bigfork chert at Black Knob Ridge: (1) bedded cherty shales (Pl. 1, fig. 1), (2) siliceous limestone and limestone-replacement chert (Pl. 1, figs. 2, 3), and (3) nodular and concentric chert. The bedded cherty shales are black, brown, and gray. In many places they are irregularly laminated, wavy bedded, color-banded, pyritiferous, and sapropelic. Many of the rocks contain relict carbonate rhombs. Most of the cherty shales are fractured and veined with quartz and/or calcite. In some specimens, the "argillaceous" impurities are

finely divided organic matter rather than clay minerals. Cryptocrystalline silica predominates in the light-colored layers, whereas the silica in the dark sapropelic layers is more nearly isotropic. The change from light to dark laminae is very abrupt in some specimens and is conspicuous even in hand specimens. Spores (?), poorly preserved Radiolaria, and a few sponge spicules are the most abundant fossils in the lower bedded cherty shales (Pl. 1, fig. 1). These bedded cherty shales were probably predominantly siliceous from their time of deposition, although the relict carbonate rhombs indicate that some carbonate was present originally.

The second major lithologic type in the lower unit of the Bigfork chert is siliceous limestone. There is a complete series ranging from nearly pure limestone (Pl. 1, fig. 2) through siliceous limestone to limestone-replacement chert in which only traces of carbonate remain (Pl. 1, fig. 3). The limestones range in particle size from medium crystalline to sublithographic. Dolomite, where present, is in small euhedral to subhedral rhombs. The silica in the siliceous limestones is mostly cryptocrystalline, but there is some chalcedonic silica. Small amounts of clay minerals and pyrite are present in many specimens. Sponge spicules, ostracod carapaces, and graptolite fragments are abundant in many of these limestones; crinoid columnals and fragmentary fossils of other types are less common. Most of the sponge spicules are partially silicified calcareous sponge spicules, but some appear to have been siliceous originally.

The nodular chert and concentric chert consist of very siliceous, very finely crystalline to sublithographic limestone and light-colored calcareous chert. These cherts consist of apparently random mosaics of light-yellow "isotropic" chert, subhedral to euhedral carbonate rhombs, and segments of calcareous to partially silicified sponge spicules. Small amounts of brownish sapropelic matter, pyrite, and chitinous (?) fragments of fossils are present. Quartz veinlets occur in some specimens. The chert and shale of the relatively thin upper unit of the Bigfork at Black Knob Ridge strongly suggest deposition under reducing conditions in stagnant water unfavorable to the existence of most bottom-dwelling organisms (euxinic

Plate 1. PHOTOMICROGRAPHS

Figure 1. Color-Banded Bigfork Cherty Shale, Black Knob Ridge

Thinly laminated, color-banded, sapropelic cherty shale. The division between the dark-colored sapropelic layers containing "isotropic" silica and the light-colored layers with cryptocrystalline silica is sharply defined. The specimen contains scattered carbonate rhombs and is cut by a network of veinlets, some of which contain carbonate. Thin Section BF-215, plane light, × 6.3.

Figure 2. Ostracodal Bigfork Limestone, Black Knob Ridge

Slightly siliceous, medium crystalline, ostracodal limestone. There is a strong development of twinning and gliding lines in the calcite crystals. At a few places the rock is cut by veinlets of coarsely crystalline clear calcite. There are small amounts of pyrite and asphaltic (?) material. BF-242, plane light, × 65.

Figure 3. Calcareous Bigfork Chert, Black Knob Ridge

Fractured and veined, calcareous, limestone-replacement chert. Veinlets and stringers of calcite are abundant. The dark-colored areas consist of an amorphous clay-silica mixture which is white under reflected light. Ostracod carapaces, sponge spicules, and crinoid columnals are scattered through the rock; some of them are pyritized. BF-13, plane light, × 5.5.

Figure 4. Color-Banded Chalcedonic Bigfork Chert, Black Knob Ridge

Color-banded chert from the upper unit of the Bigfork chert. This specimen consists of alternating layers of dark red-brown sapropelic chert, cryptocrystalline silica, and clear chalcedony. At places in the chalcedonic layers there are mosaics of interlocked silt-size quartz. Fragmentary graptolites and poorly preserved radiolaria are abundant in the dark-colored layers. BF-260, crossed nicols, × 28.

Figure 5. Silty Chloritic Graphitic Bigfork Chert, McCurtain County

Silty graphitic chert. The clay minerals have been reconstituted into chlorite and sericite. The graphitic bands have been folded and offset by microfaults, thus giving the specimen a pseudo-brecciated appearance. BF-127, plane light, × 17.

Figure 6. Siliceous Chlorite Argillite, Bigfork Chert, McCurtain County

Siliceous chlorite argillite. Abundant thin streaks and aggregates of chlorite and sericite and small amounts of silt-size quartz are disseminated through a matrix of cryptocrystalline silica. BF-128, crossed nicols, × 65.

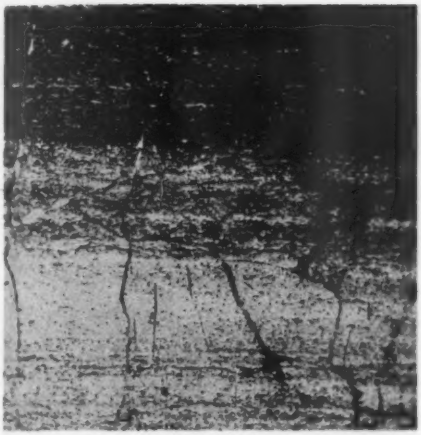

FIGURE 1

FIGURE 2

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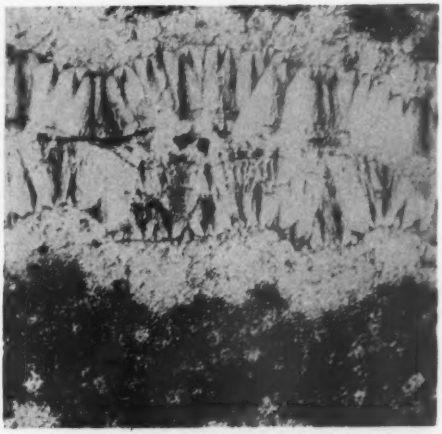

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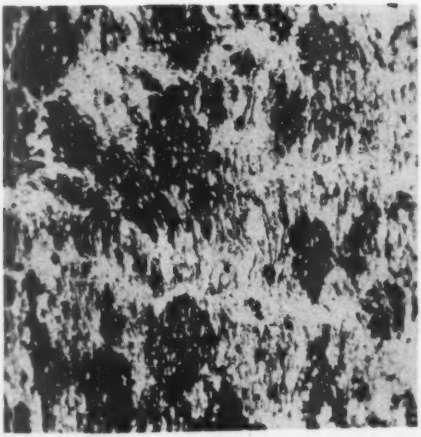

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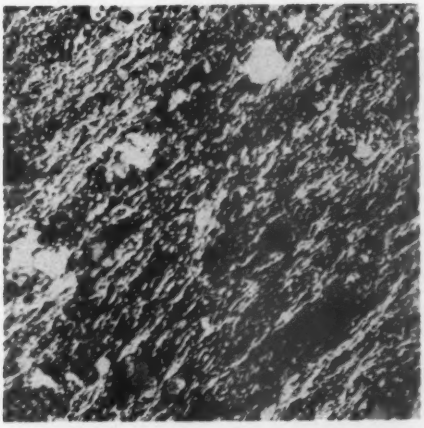
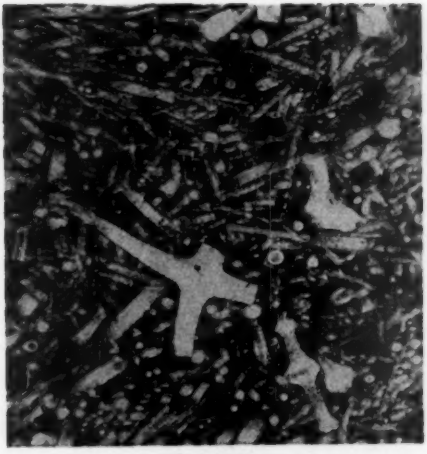


FIGURE 6

PHOTOMICROGRAPHS



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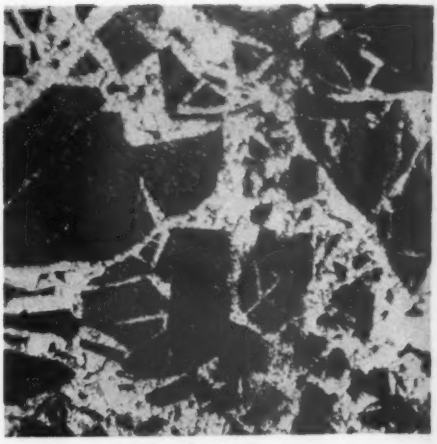

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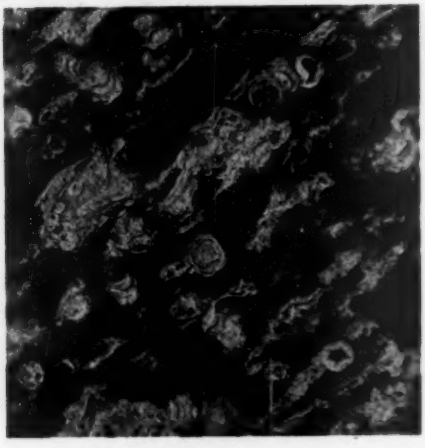

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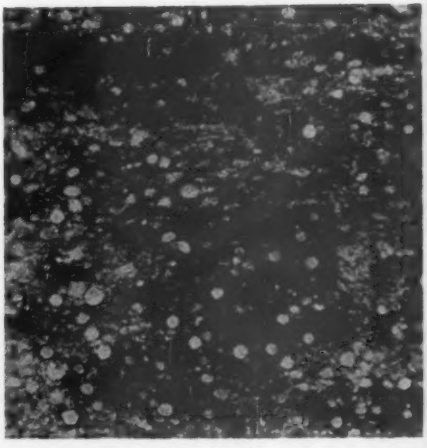

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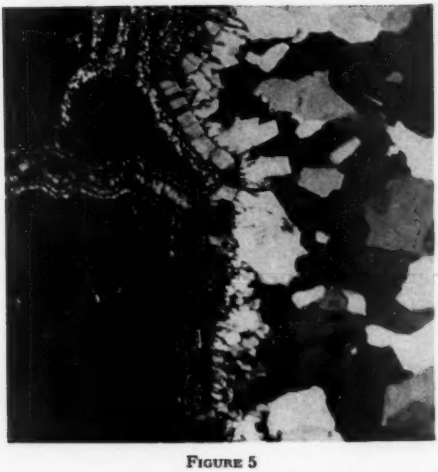

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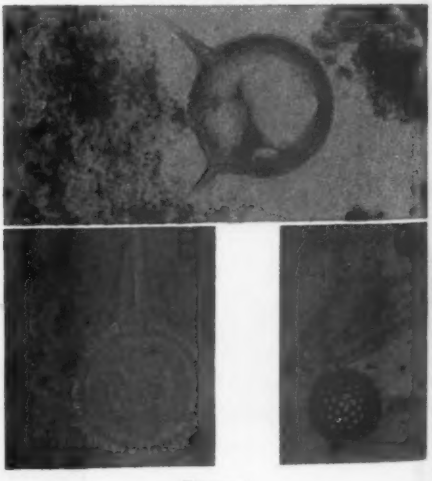

FIGURE 6

PHOTOMICROGRAPHS

environment). No calcareous fossils and very little carbonate are present. The chert is dark red-brown, laminated, pyritiferous, and sapropelic. It contains spores, Radiolaria, and sponge spicules. Veinlets or fissure fillings of black asphaltic (?) material are noted in many specimens. Most slides cut perpendicular to the bedding show thin, irregular, curving, subparallel streaks of dark opaque carbonaceous and bituminous material and red-brown organic detritus in a matrix of "isotropic" silica, pyrite, and small amounts of clay minerals. These rocks are sharply color-banded at places. The contact between the dark and light layers is abrupt and is marked at most places by a change from "isotropic" silica to chalcedonic or cryptocrystalline silica (Pl. 1, fig. 4).

The black shale irregularly interbedded with the sapropelic chert contains less silica and more organic matter and argillaceous material than the chert.

McCurtain County, Oklahoma

Stratigraphy.—The Bigfork chert in McCurtain County, Oklahoma, has been described by Honess (1923) as a series of uniformly bedded black cherts, coal-black shales, and black cherty limestones. It resembles closely the Bigfork chert in western Arkansas which has been described in considerable detail by Miser and Purdue (1929). They state (p. 37) that:

"The formation, . . . , consists of chert interbedded with some shale and limestone. The chert is commonly in even-bedded layers having a thickness of 3 to 6 inches but at a few places thickening to 3 feet. It is very close textured and brittle and under the blows of a hammer flies into small pieces, most of which have an uneven but some a conchoidal fracture. Parts of it are finely laminated. It is black and dense when unweathered, but when weathered it ranges in color from slate to dark gray, the slate color occurring at most places. Some of the chert is rather porous and has the appearance of gray, soft, fine-grained sandstone. In some parts small quantities of calcite and pyrite are disseminated. Numerous joints, many of which are straight and have remarkably smooth glossy surfaces, form a network which cuts the layers in all directions. So numerous are the joints that it is difficult to obtain a hand specimen with fresh surfaces. In most parts the fractures along the joints are occupied by fine quartz veins containing a little calcite. The layers as a rule are greatly crumpled, and it probably was the strain accompanying the crumpling that formed the network of joints.

PLATE 2. PHOTOMICROGRAPHS

FIGURE 1. SPICULITE CHERT, ARKANSAS NOVACULITE, BLACK KNOB RIDGE

Fractured and veined spiculite chert from the lower member. It consists of sponge spicules of many different sizes and shapes embedded in a siliceous matrix. The matrix in this specimen consists of "isotropic" silica which is brown in transmitted light and white in reflected light, but at places the silica is cryptocrystalline. Locally, the spicules may be recrystallized and their outlines are poorly defined. AN-10, plane light, × 69.

FIGURE 2. BRECCIATED ARKANSAS NOVACULITE, POTATO HILLS

Brecciated Arkansas novaculite from the lower middle member. It consists of light brown "isotropic" chert which has been brecciated and re-cemented with vein quartz and cryptocrystalline silica. AN-275, crossed nicols, × 28.

FIGURE 3. SPORE-BEARING SAPROPELIC CHERT, ARKANSAS NOVACULITE, BLACK KNOB RIDGE

Chalcedonic, pyritiferous, spore-bearing sapropelic chert from the lower middle member. The rock is thinly and irregularly laminated. Pyrite is disseminated widely through the dark-colored sapropelic laminae. Most of the numerous spore exines are silicified and somewhat deformed. AN-36B, plane light, × 65.

FIGURE 4. RADIOLARIAN SILICEOUS SHALE, ARKANSAS NOVACULITE, BLACK KNOB RIDGE

Green, chloritic, radiolarian siliceous shale from the upper middle member. The radiolaria are very abundant but generally are not as well-preserved as in some of the cherts. AN-254, plane light, × 29.

FIGURE 5. COLLOFORM STRUCTURE, ARKANSAS NOVACULITE, BLACK KNOB RIDGE

Quartz-chalcedony "geode" with colloform structure from the upper member. This photomicrograph shows only a portion of the "geode". The dark area at left is brown chert. The center of the geode consists of a mosaic of quartz which grades into brown feathery chalcedony and then into cryptocrystalline silica at the outer edge. Many specimens have colloform veinlets from the "geode" into the surrounding chert similar to those shown in the photomicrograph. AN-37, crossed nicols, × 29.

FIGURE 6. RADIOLARIA IN ARKANSAS NOVACULITE, BLACK KNOB RIDGE

Well-preserved radiolarian tests from upper member. Specimen C is pyritized.

A. AN-248, plane light, × 121.

B. AN-28, plane light, × 129.

C. AN-42, plane light, × 88.

The shale is black, siliceous, and carbonaceous and forms even-bedded layers ranging from a fraction of an inch to several feet in thickness which are distributed throughout the formation, but is meager or absent in places, particularly near the base. Black siliceous limestone in lenses and layers only a few inches thick occurs sporadically near the base and top."

Petrography.—The generally poor exposures, complicated folding, and lack of measured sections of the Bigfork chert in McCurtain County preclude any attempt to subdivide this formation into members. The Bigfork chert is somewhat more metamorphosed here than at Black Knob Ridge, but the grade of metamorphism attained is comparatively low.

Near the lower (Womble) contact, the Bigfork consists of fractured and veined microcrystalline chert (Pl. 1, fig. 5). At places the specimens are ferruginous and graphitic. The ferruginous material and finely dispersed black (graphitic?) opaque dust occurs either in irregular blotches or in wavy discontinuous streaks parallel to the apparent bedding. The quartz veinlets cutting the rock are typically contorted, almost ptygmatic in their folding. Most of the fossils in these cherts are quite fragmentary; crinoid columnals, brachiopods, ostracods, and sponge spicules are present. The bulk of the lower and middle Bigfork consists of similar limestone-replacement cherts with marine fossils, but there are some impure, pyritiferous, graphitic cherts containing a few spores and Radiolaria. Strong cataclastic effects are evident in many specimens.

The upper Bigfork differs lithologically to some degree from the lower and middle Bigfork. The major lithologic types, chloritic chert and silty chlorite argillite, grade into each other. The cherts are veined and contorted. In some slides there are two well defined directions of alignment of the long axes of the micas and chlorite. The chloritic argillites are silty in many cases; at places they grade into chloritic siltstones. The aligned wisps and shreds of chlorite, sericite, silt-size quartz, and clay minerals are embedded in a matrix of cryptocrystalline silica (Pl. 1, fig. 6). There are highly contorted and twisted streaks of mica and chlorite in some specimens; false cleavage and incipient phyllitization are shown in some slides.

Although exact correlation is not possible, the fossiliferous chert of the lower and middle Bigfork in McCurtain County is probably equivalent to the lower limestone-chert unit of the Bigfork chert at Black Knob Ridge. The chloritic argillite and chert of the upper Bigfork in McCurtain County are probably metamorphosed equivalents of the black paper shale and sapropelic chert of the upper unit of the Bigfork at Black Knob Ridge.

Arkansas Novaculite (Devonian-Mississippian)

General Statement

The Arkansas novaculite, the best-known yet most controversial of the siliceous sediments of Ouachita facies, contains a wide variety of lithologic types, including not only novaculite but also spiculite chert, calcareous chert, radiolarian chert, sapropelic chert, siliceous shale, and sapropelic paper shale. Most of the descriptions of Arkansas novaculite have been of rocks from western Arkansas and McCurtain County, Oklahoma, where it has been subjected to metamorphism. The sequence of siliceous sedimentary rocks in the Arkansas novaculite at Black Knob Ridge is less metamorphosed and more closely approximates the original lithology of the sediment.

Black Knob Ridge

Stratigraphy.—Hendricks *et al.* (1947) state that the Arkansas novaculite overlies the Missouri Mountain shale with apparently gradational contact and that the Arkansas novaculite

"... ranges in thickness from 234 to 340 feet in sections measured in Black Knob Ridge, and is made up of four lithologic members. The lowest member constitutes about half of the formation; it is made up of thin-bedded, light-colored novaculite, much of which is apple-green in color, interbedded with hard green shale. The second lithologic member constitutes about half of the upper part of the formation. It is made up of beds of novaculite of variable thickness that are light gray to almost black, and are separated by thin partings and beds of black paper shale. The basal beds of this member are generally conglomeratic. The two uppermost members are about equal in thickness and together constitute about one fourth of the formation. The lower of these two members consists of red and green, gritty, micaceous, and flaky shale that weathers to a soft clay; this shale contains some thin beds of novaculite, particularly in the lower part, and some local beds of blocky black shale. The upper member consists of green, brown, and gray novaculite in beds ranging from 1 inch to 1 foot thick."

The fauna and flora of the novaculite consists of sponge spicules, radiolarian capsules, microspores, collapsed megaspores, probable charophyte oogonia, silicified wood, linguloid brachiopods, and conodonts. Recent studies of the conodont fauna by W. H. Hass (1951) show that sediments of Devonian and Mississippian age are present in the Arkansas novaculite.

Petrography.—The *lowest member* of the Arkansas novaculite at Black Knob Ridge consists predominantly of massive spiculitic chert and laminated siliceous shale or argillite. The "matrix" of the chert is composed largely of cryptocrystalline silica with minor amounts of "isotropic" silica and microcrystalline silica. Quartz veinlets, fractures, and joints appear in about half of the specimens but few of the rocks can be described as brecciated. Clays, hydrous iron oxides, and "floated" sand grains are the most abundant impurities. About half of the specimens contain some sericite, chlorite, chalcedony, and bituminous (?) material. Glauconite, colloform structures, and color-banding are found in a few slides.

Radiolaria are scarce but sponge spicules are abundant. Some specimens are composed almost entirely of sponge spicules (Pl. 2, fig. 1). In the cherts containing some argillaceous material, the spicules are outlined sharply in most cases; in the white, non-argillaceous cherts, the spicules tend to be "resorbed" into the cryptocrystalline matrix and are difficult to see clearly.

The *lower middle member* of the Arkansas novaculite at Black Knob Ridge consists of (1) dark, sapropelic, spore-bearing chert, and black paper shale; (2) light-colored, argillaceous, radiolarian chert; and (3) radiolarian siliceous shale. The sapropelic chert and black shale are laminated in most cases and contain abundant red-brown organic matter, pyrite, and bitumen; they suggest deposition in an euxinic environment. The silica in the lighter-colored laminae is predominantly cryptocrystalline. Much of the silica in the dark laminae of the sapropelic cherts is apparently isotropic, but the index of refraction of this silica is higher than that of opal. This "isotropic" silica is apparently identical with that in the Bigfork chert at Black Knob Ridge. Similar non-opaline isotropic silica in cherts has been reported by Lawson (1895), Davis (1918, p. 256), and Sargent (1929, p. 402). In the Arkansas novaculite, this "isotropic" silica apparently has an index of refraction slightly higher than Canada balsam (1.54). The light-colored cherts and siliceous shales contain cryptocrystalline silica. Many of the dark sapropelic cherts are thinly laminated, whereas the light-colored cherts and siliceous shales are rarely thinly laminated. Almost every specimen from this member is fractured to some extent and contains fissure fillings of quartz; a few of the specimens have been brecciated and recemented with quartz (Pl. 2, fig. 2). "Floated" sand grains, micas, chlorite, chalcedony, and glauconite are uncommon.

Very few sponge spicules are found in the lower middle member. In contrast, the light-colored cherts and siliceous shales contain abundant well-preserved radiolarian capsules at places. The dark sapropelic cherts generally contain abundant spore exines (Pl. 2, fig. 3); most are well-preserved microspores and are oval to spherical in cross-section, but some are collapsed and flattened megaspores.

The *upper middle member* of the Arkansas novaculite at Black Knob Ridge is apparently transitional between the lower middle and upper members and contains a variety of lithologic types. The conspicuous red and green "shales" which dominate this member range from radiolarian shale (Pl. 2, fig. 4) to siliceous and chloritic argillite. The chert and novaculite includes white cryptocrystalline radiolarian chert, argillaceous chert, and dark sapropelic chert containing both Radiolaria and spores. Sponge spicules are uncommon. The more competent beds of chert contain quartz veinlets and show minor fracturing, whereas the shales and argillites are seldom fractured or veined.

Almost every specimen from the *upper member* contains abundant Radiolaria (Pl. 2, fig. 6). Most of the siliceous rocks from this member are cryptocrystalline argillaceous radiolarian chert, radiolarian shale, and argillite. About half of the specimens examined are laminated. Almost all of the slides show some fracturing and subsequent fissure-filling by quartz. Hydrous iron oxide is widespread. Micas and chlorite are abundant in the shale and argillite. Quartz geodes, colloform structure (Pl. 2, fig. 5), and color-banding are present in some slides.

Near the contact with the overlying Stanley

formation, there are at places a few beds of very fine-grained subgraywacke sandstone and siltstone, which closely resemble similar sandstone in the lower Stanley but are here overlain by typical novaculite. The actual contact is difficult to pick at some localities.

McCurtain County, Oklahoma

Stratigraphy.—Honess (1923, p. 110) considers that the Arkansas novaculite in McCurtain County, Oklahoma, is practically identical lithologically with the Arkansas novaculite in the DeQueen and Caddo Gap quadrangles, Arkansas. Honess differs from Miser's (1917) description of the stratigraphy only in considering the carbonate mineral in the upper member of the novaculite manganese carbonate rather than manganese-bearing calcite.

Miser and Purdue (1929, p. 50–53) described fully the lithology of the Arkansas novaculite in western Arkansas. Part of their description is as follows:

"The formation, as developed in these quadrangles and in many other parts of the Ouachita Mountains, consist of three lithologic divisions—a lower one, made up almost entirely of massive, white novaculite; a middle one, consisting mainly of thin layers of dense, dark-colored novaculite interbedded with shale; and an upper one, consisting chiefly of massive, highly calcareous novaculite. These divisions differ in thickness and character from place to place."

Petrography.—The petrography of the Arkansas novaculite in McCurtain County has been described in considerable detail by Honess (1923, p. 109–139). In general the novaculite in McCurtain County differs considerably in fabric from the novaculite at Black Knob Ridge but is similar in composition. Most of the rock types described from Black Knob Ridge are recognizable in McCurtain County and the general sequence of sediments is similar. Some major differences are:

1. Cataclastic effects are more widespread and more conspicuous in the McCurtain County novaculite. Microscopic "augen," schistose and granulated zones, brecciated areas, contorted quartz veinlets, and microstylolites are developed in many specimens.

2. Siliceous fossils are deformed and tend to be "resorbed" into the cryptocrystalline silica matrix of the rock so that their outlines are difficult to see in transmitted light. In some cases siliceous sponge spicules are so incorporated into the matrix that they can be seen only in the argillaceous portions of the rock or where they are outlined by opaque dust-size particles. The radiolarian capsules are deformed in many specimens; generally they are compressed and flattened and have lost most of their spines (Pl. 3, fig. 1). The Radiolaria are recrystallized in some cases into a clear, lenticular area of silt-size quartz with no visible internal structure.

3. The grain size and crystallinity of the siliceous matrix is increased. Much of the silica in the Arkansas novaculite at Black Knob Ridge is in the "isotropic" and chalcedonic state and the remainder is cryptocrystalline. In McCurtain County, isotropic and chalcedonic silica are exceedingly rare, and much of the silica is "microcrystalline", in which the individual particles are within the silt range on the Wentworth scale.

4. There is a convergence of lithology in the McCurtain County novaculite which is the result of dynamic metamorphism. The original sedimentary textures are obscured and at places eliminated by the deformation and recrystallization, so that the rocks are generally more similar in appearance and petrographic character. Most of the argillaceous material in these rocks has been reconstituted into sericite and chlorite. Much of the sapropelic and carbonaceous material has been altered to finely disseminated graphitic material.

Several lithologic types in the specimens examined from McCurtain County were not observed in the specimens from Black Knob Ridge. The first is a widespread microbreccia or microconglomerate which consists of granule and sand-size particles of various types of chert, siliceous shale, and "floated" sand and silt grains (Pl. 3, fig. 2) and does not appear to be localized at any horizon.

The other new lithologic types are found in the upper member of the Arkansas novaculite in McCurtain County. This member consists largely of light-colored calcareous novaculite and is not similar to any of the four members at Black Knob Ridge. Generally this calcareous novaculite contains many small subhedral to euhedral rhombs of manganiferous carbonate. This same type of chert is found also in the

Walnut Creek section of the Potato Hills (Pl. 3, fig. 3). At one locality in Beaver Bend State Park in McCurtain County, the calcareous novaculite contains abundant partially silicified calcareous sponge spicules (Pl. 3, fig. 4). The carbonate weathers faster than the siliceous material and a spongy shapeless porous mass of silica is noted in most exposures.

Upper Paleozoic Siliceous Shales

Stratigraphy

The stratigraphy of the siliceous shales and adjacent beds of the Stanley, Jackfork, and Atoka formations has been discussed in detail by Harlton (1938). These siliceous shales are more resistant than most of the adjacent strata and tend to weather out in relief. Although comparatively thin, the siliceous shales maintain a uniform appearance over distances of many miles and are useful marker beds in mapping surface structures of the Ouachita Mountains in Oklahoma. The names used in describing these beds are those originated by Harlton (1938), but the names have been restricted to apply to the siliceous shale beds only in those instances where Harlton applied the name to the siliceous shale and overlying sandstone and shale beds. Also the siliceous shale beds are considered members of the Stanley, Jackfork, and Atoka formations rather than as members of smaller formations with the Stanley and Jackfork considered groups.

Petrography

Stanley shale (*Mississippian*).—Four siliceous shale members have been recognized in the Stanley formation. From oldest to youngest these are (1) Basal Ten Mile Creek siliceous shale, (2) a previously unnamed siliceous shale outcropping near the Potato Hills which is herein named the Tuskahoma siliceous shale, (3) Middle Ten Mile Creek siliceous shale, and (4) Moyers siliceous shale.

BASAL TEN MILE CREEK SILICEOUS SHALE: This siliceous shale is typically dark and is seldom conspicuously laminated. At places it is silty, micaceous, and ferruginous. The silica is evenly disseminated and almost isotropic. Harlton (1938) has referred to this siliceous material as weathered volcanic glass. Small amounts of carbonaceous material and micaceous minerals are noted in the groundmass. There are a few poorly preserved spores and recrystallized Radiolaria.

TUSKAHOMA SILICEOUS SHALE NEAR THE POTATO HILLS: The siliceous shale of this unit has not been named previously and is herein named the Tuskahoma siliceous shale. The type locality is selected as the exposure at the center of the west side of Section 22, Township 2 North, Range 20 East, Pushmataha Co., Oklahoma, approximately 4.6 miles east of Tuskahoma on U. S. Highway 271. The type section was measured by T. A. Hendricks and Paul Averitt (1947, p. 35). This shale is typically carbonaceous and pyritiferous. Some specimens contain abundant sapropelic and/or carbonaceous material, micaceous minerals, iron oxides, and quartz veinlets. Many shales from this unit are conspicuously laminated; thin sections perpendicular to bedding planes show thin, wavy, crenulated, discontinuous lenses of clay minerals and sapropelic material intermingled with cryptocrystalline to "isotropic" silica. Spore exines are the most characteristic fossils. Delicate radiolarian structures are well-preserved in a few specimens, and scattered sponge spicules and radiolarian spines are not rare.

MIDDLE TEN MILE CREEK SILICEOUS SHALE: This unit consists largely of mottled dark siliceous shale which is micaceous, pyritiferous, and carbonaceous. Many specimens are laminated; individual laminae are thin and irregular. Quartz veinlets and red-brown sapropelic matter are present in some thin sections. Radiolaria are abundant but poorly preserved. Only a few spores (?) and sponge spicules were noted.

MOYERS SILICEOUS SHALE: This dark siliceous shale is generally thinly laminated and irregularly banded. Individual laminae are wavy, crenulated, and discontinuous. The disseminated silica is isotropic for the most part, but there are a number of streaked, crumpled, shapeless masses of cryptocrystalline silica. Embedded in the matrix of clay minerals, carbonaceous material, and silica are some remarkably well-preserved radiolaria, a few spores (?), some silt-size quartz, and some small sponge spicules (both monaxons and hexaxons).

Jackfork sandstone (*Pennsylvanian*).—Four

siliceous shales have been recognized in the Jackfork formation. From oldest to youngest these are (1) Chickasaw Creek siliceous shale, (2) Prairie Mountain siliceous shale, (3) Markham Mill siliceous shale, and (4) Wesley siliceous shale. Both the Chickasaw Creek and Wesley siliceous shales contain lenses of chert conglomerate.

CHICKASAW CREEK SILICEOUS SHALE: This unit consists largely of brown radiolarian siliceous shale. The shales are generally irregularly laminated and wavy bedded. The matrix of the rock consists of "isotropic" silica, sapropelic or carbonaceous material, and clay minerals. There are quartz veinlets in a few specimens; at places the clay minerals adjacent to the quartz veinlets are reconstituted to micas. There are small amounts of pyrite, silt-size quartz, and micas. Spores (?) are not abundant, but sponge spicules and radiolaria are very abundant in some specimens. Most of the radiolaria are concentrated in the light-colored laminae.

Lenses of conglomerate or breccia as much as 18 inches thick occur in the Chickasaw Creek siliceous shale. The lenses are not uniform lithologically, but consist largely of brecciated pebbles and granules of chert and siliceous shale cemented by chalcedonic chert (Pl. 3, fig. 5). Quartz sand, pyrite, epidote, hematite, and minor amounts of other minerals are included. Many quartz veinlets cut through the rock; in some specimens these veinlets show fine-grained margins and comb structure.

PRAIRIE MOUNTAIN SILICEOUS SHALE: This unit consists largely of dark siliceous shale which is conspicuously banded and thinly but irregularly laminated. Most of the silica is isotropic and the thin sections appear almost completely dark under crossed nicols. Tiny

PLATE 3. PHOTOMICROGRAPHS

FIGURE 1. DEFORMED RADIOLARIA, ARKANSAS NOVACULITE, MCCURTAIN COUNTY, OKLAHOMA

Pyritiferous, graphitic (?), cryptocrystalline chert with abundant deformed radiolaria. Most of the spines are broken; only those parallel to the direction of compression of the rock remain. AN-153, plane light, × 65.

FIGURE 2. INTRAFORMATIONAL CHERT BRECCIA, ARKANSAS NOVACULITE, MCCURTAIN COUNTY, OKLAHOMA

Graphitic (?), pyritiferous, chert breccia. The angular to subangular, granule to sand-size fragments of spiculite chert, radiolarian chert, massive chert, and siliceous shale are embedded in a matrix of microcrystalline silica. The rock is cut by numerous quartz veins, some of which are highly contorted. AN-103, plane light, × 5.7.

FIGURE 3. CALCAREOUS UPPER MEMBER OF ARKANSAS NOVACULITE, POTATO HILLS, PUSHMATAHA COUNTY, OKLAHOMA

Veined and fractured chert containing numerous subhedral to euhedral carbonate rhombs. The matrix in which the carbonate is embedded consists of equigranular cryptocrystalline silica. The carbonate rhombs vary in size from sublithographic to finely crystalline. AN-40, plane light, × 65.

FIGURE 4. CALCAREOUS SPICULITE CHERT FROM UPPER MEMBER OF ARKANSAS NOVACULITE, MCCURTAIN COUNTY, OKLAHOMA

Calcareous chert consisting largely of sponge spicules which were originally calcareous. Most of the sponge spicules are partially silicified in this specimen. The spicules are embedded in a matrix of cryptocrystalline to microcrystalline silica. The rock is cut by numerous sinuous discontinuous shear planes. There are a few detrital grains of silt-size quartz. AN-171, plane light, × 58.

FIGURE 5. CHICKASAW CREEK CHERT BRECCIA

Unsorted, granule to pebble-size chert breccia. This rock is from a lens in the Chickasaw Creek siliceous shale member of the Jackfork sandstone. It consists of fragments of siliceous shale, dark red-brown sapropelic chert, and light-colored argillaceous chert embedded in a matrix of chalcedony and quartz. Abundant cubes and anhedral aggregates of pyrite are scattered through the matrix. Veins of quartz cut through the pebbles and granules; most of these veins are simple fissure-fillings but some show comb structure and fine-grained margins. There are a few radiolaria in the matrix. JF-239, plane light, × 5.7.

FIGURE 6. MARKHAM MILLS SILICEOUS SHALE

Laminated, sandy, fossiliferous siliceous shale from the Markham Mills member of the Jackfork sandstone. Thin bands of poorly sorted, quartzitic, glauconitic coarse siltstone and very fine-grained sandstone alternate irregularly with laminae of carbonaceous siliceous shale. Sponge spicules are very abundant in some layers and there are some radiolaria and spore exines. H-31, plane light, × 24.

FIGURE 1

FIGURE 2

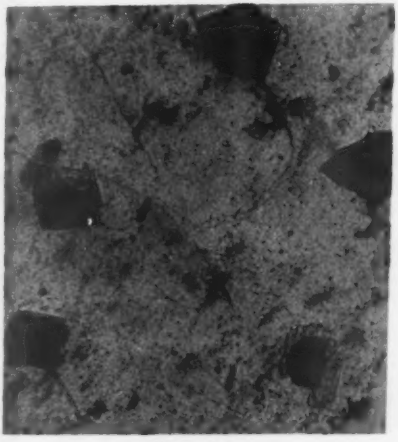

FIGURE 3

FIGURE 4

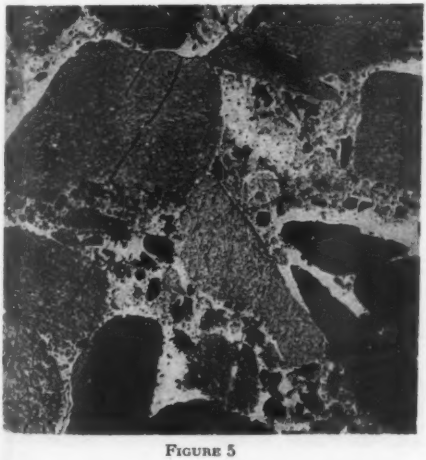

FIGURE 5

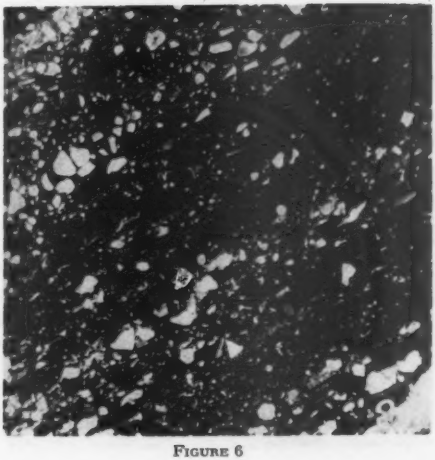

FIGURE 6

PHOTOMICROGRAPHS

FIGURE 1

FIGURE 2

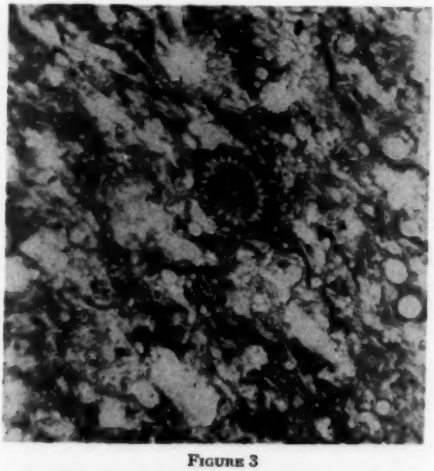

FIGURE 3

FIGURE 4

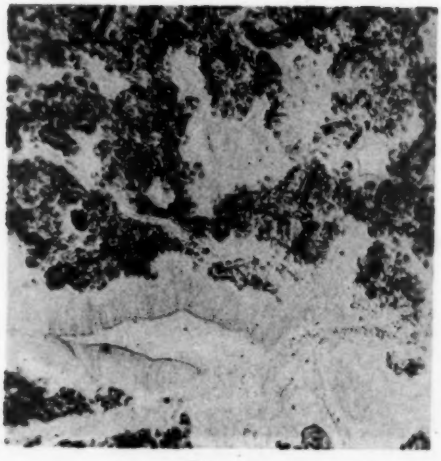

FIGURE 5

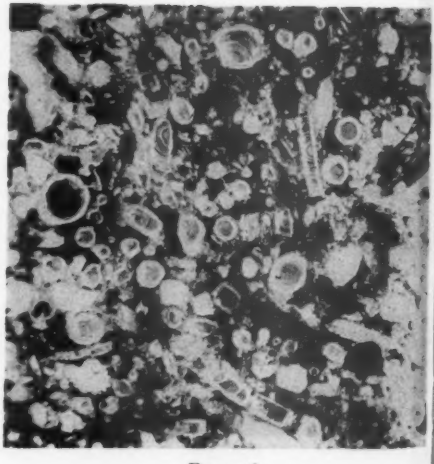

FIGURE 6

PHOTOMICROGRAPHS

flakes of mica scattered through the shale extinguish parallel to bedding planes. Hydrous iron oxide is relatively abundant; the conspicuous banding results from changes in the relative amounts of clay and iron oxides. Radiolaria and spores are abundant, but there are only a few sponge spicules.

MARKHAM MILL SILICEOUS SHALE: This unit contains strongly laminated, lithologically complex shales (Pl. 3, fig. 6). The silica is almost entirely cryptocrystalline. Some specimens consist of thin, irregular, discontinuous layers of micaceous siliceous argillite interbedded with relatively clear lenses and layers of chert containing radiolaria, spores, sponge spicules, and abundant tiny spicular bodies (radiolaria spines?). Other specimens show rapid gradations from spiculitic, carbonaceous, radiolaria, and spore-bearing siliceous shale to poorly sorted glauconitic siltstone with silicified crinoid columnals, giving the overall impression of rapid alternations of stagnant and normal marine environments during deposition.

WESLEY SILICEOUS SHALE: The Wesley siliceous shale unit (Pl. 4, figs. 1, 2) is somewhat variable lithologically. It consists largely of spiculitic chert, spiculitic and chloritic siliceous argillite, and dark carbonaceous siliceous shale. Some of the shale is laminated, with typical microscopic crenulations and twistings of each thin layer of carbonaceous material, clay, and silica. Sponge spicules predominate over radiolaria and spores in most specimens; one specimen contains abundant collapsed and flattened megaspores. Fragments of fossils are much more plentiful in the light-colored layers than in the dark layers; the light-colored layers are also harder and more siliceous.

Lenses of chert conglomerate are intercalated in the Wesley siliceous shale. These rocks are

PLATE 4. PHOTOMICROGRAPHS

FIGURE 1. WESLEY SILICEOUS SHALE

Spiculitic siliceous shale from the Wesley member of the Jackfork sandstone. Thin section cut parallel to the bedding plane. Cruciform and monaxon sponge spicules are embedded in a matrix of clay minerals, silica, and finely disseminated hematite "dust". H-28, plane light, × 65.

FIGURE 2. WESLEY SILICEOUS SHALE

From the same specimen as in Figure 1, but cut perpendicular to the bedding plane. The crenulated, discontinuous nature of individual laminae of argillaceous and organic material may be seen. Part of the argillaceous material is reconstituted into chlorite. Clear lens-shaped areas of chert are scattered through the rock. Many of the sponge spicules are seen in cross-section in this photomicrograph and appear as small circles. H-30, plane light, × 65.

FIGURE 3. LOWER ATOKA SILICEOUS SHALE

Sandy, spiculitic, siliceous shale from the Atoka formation. Cut perpendicular to the bedding plane. The wavy, discontinuous streaks and layers of argillaceous material and carbonaceous material are surrounded by a matrix of cryptocrystalline silica. Sponge spicules and detrital quartz grains are distributed abundantly through the rock. There are a few radiolaria. H-46, plane light, × 64.

FIGURE 4. HATTON TUFF LENTIL

Chloritic, devitrified, silicified tuff from the Hatton tuff lentil of the Stanley shale. Vitroclastic structure is well-developed. The matrix consists largely of chlorite and cryptocrystalline silica. The shards and bubble-wall fragments appear well-preserved in plane light, but under crossed nicols they appear as a shapeless mosaic of quartz and chert. STH-1, plane light, × 65.

FIGURE 5. PINETOP CHERT

Calcareous, cryptocrystalline to chalcedonic chert from the Pinetop chert near the upper (Woodford) contact. Much of the chert is of the "vesicular", limestone-replacement type; the silica in these areas appears brown in transmitted light and white in reflected light. Colloform nodules and veinlets of chalcedony are conspicuous; the one in the lower part of the photomicrograph has a center of clear quartz surrounded by a rim of brown chalcedony. PT-5, plane light, × 65.

FIGURE 6. CHICKACHOC SPICULITE CHERT

Silty, calcareous, spiculite chert from the Chickachoc chert. This specimen is variable in lithology within a single thin section, containing areas of pure spiculite, calcareous spiculite chert, siliceous limestone, and calcareous siltstone. The sponge spicules appear to have been calcareous originally; upon silicification the outer walls of the spicules were replaced by clear chalcedony and the pore canals (inner portion) of the spicules were replaced by brown chalcedony with concentric ring structure. CC-201, plane light, × 65.

pebble to granule chert conglomerates cemented by quartz, chalcedony, and an unidentified light yellow-green isotropic mineral. The coarser fragments contain a wide variety of lithologic types, including partly devitrified tuff, "isotropic" radiolarian chert, argillaceous chert, and siliceous shale.

Atoka formation (*Pennsylvanian*).—The only siliceous shale recognized so far in the Atoka formation is in the lower part (Pl. 14, fig. 3). It consists of dark carbonaceous, spiculitic siliceous shale and spiculitic chert. The irregularly alternating layers of clay, pyritic carbonaceous material, and silica are wavy and discontinuous. Some of the rocks are somewhat silty. There are many well-preserved Radiolaria in this unit, but they are quantitatively subordinate to sponge spicules.

Hatton Tuff Lentil (Mississippian)

General Statement

Properly speaking, the Hatton tuff lentil of the Stanley shale is not one of the siliceous sediments of Ouachita facies. However, it is of the utmost significance in any discussion of the origin of the siliceous sediments, and so is briefly discussed here.

Stratigraphy

The Hatton tuff lentil of the Stanley formation is not found in the western part of the Ouachita Mountains of Oklahoma, but it is widely distributed in McCurtain County, Oklahoma, and in western Arkansas. As described by Miser and Purdue (1929), the Hatton tuff lentil is the lowest and thickest of a number of tuff beds of similar lithologic character in the lower Stanley. According to them, the Hatton tuff lentil in the DeQueen and Caddo Gap quadrangles, Arkansas, ranges from 30 to 90 feet thick, whereas the thinner tuff beds are from two to four in number and range from 6 to 40 feet thick. Honess (1923), on the other hand, doubts that there is more than one bed of the tuff in McCurtain County.

Megascopically, according to Miser and Purdue (1929, p. 64):

"All the tuffs, including the Hatton tuff, are similar in lithologic features. They are so massive and homogeneous and their joints are so well developed that if bedding planes exist they are, in most if not all places, confused with the joints, but the bedding can be determined by the parallel arrangement of numerous green, flattened pellets of chloritic material that are usually present in some parts of the beds. The color of the unweathered rock is dark gray with a greenish tinge, but that of the weathered rock, which consists of a layer usually less than an inch thick over the surface of the exposed parts, range from light greenish gray on the surface to brown next to the fresh portion. The rock is tough and compact and like other beds in the Stanley contains quartz veins, but the veins seem more numerous in the Hatton lentil than elsewhere in the Stanley. In its macroscopic character it strikingly resembles a porphyritic igneous rock. The largest conspicuous grains are feldspars, which are usually 2 millimeters or less in diameter, but some are as much as 5 to 7 millimeters. (. . .) On weathering, the grains of feldspar kaolinize and become white, giving the rock a spotted appearance."

Petrography

Miser and Purdue (1929, p. 64) quote E. S. Larsen to the effect that the original rock was near a rhyolite or quartz latite.

Honess (1923, p. 179–188) has discussed the microscopic petrography of the tuffs near the base of the Stanley in McCurtain County in some detail. He describes a number of specimens and presents a chemical analysis and a calculated norm. These analyses indicate that the tuffs are acidic.

Messrs. C. A. Weintz and A. Goldstein, Jr. collected several specimens from the type locality of the Hatton tuff lentil half a mile south of Hatton, Arkansas (Pl. 4, fig. 4). The specimens consist of greenish, chloritic, devitrified, silicified crystal-vitric tuff with well-developed vitroclastic structure (Bogenstruktur). Most of the larger grains of potash feldspar, sodic plagioclase, perthite, quartz, and igneous rock fragments are pyroclastic, but some of the grains of quartz and rock fragments are probably of sedimentary origin. The outlines of the extremely angular, sharply concave and convex, originally glassy shards and bubble wall fragments are well-defined in plane-polarized light; under crossed nicols, it is noted that they have been recrystallized into a mosaic of anhedral intercrystallized quartz and chert. The matrix of the rock consists of a mixture of chlorite and cryptocrystalline silica; veins, streaks, and irregular blobs of massive chlorite are also present. There are small amounts of sericite, iron oxides, and clay minerals.

Siliceous Sediments of the Faulted Zone between the McAlester Basin and the Central Ouachita Mountains

General Statement

Successive thrust slices in the frontal zone of the Ouachita Mountains in Oklahoma contain a sequence of sedimentary rocks which more closely resemble those exposed in the Arbuckle Mountains than those of the central Ouachita Mountains (Hendricks *et al.*, 1947). The siliceous rocks of this faulted zone comprise the Pinetop chert of Middle Devonian age, the Woodford chert of Devonian (?) age, and the Chickachoc chert of Pennsylvanian age. It is considered probable that these rocks were deposited farther out towards the foreland than those of Ouachita facies typical of the central Ouachita Mountains; their intermediate character probably results from deposition in an intermediate position between sediments of Arbuckle and Ouachita facies.

Pinetop Chert (Devonian)

The Pinetop chert is exposed in the fault block southeast of the Pine Mountain fault. According to Hendricks *et al.* (1947), the Pinetop chert is about 40 feet thick, white to light gray, and contains lenses of very fine-grained, sparingly fossiliferous limestone.

Thin sections of the Pinetop chert consist of siliceous limestone, dolomitic limestone, and calcareous cryptocrystalline to chalcedonic chert. The limestones are clean to slightly argillaceous, inequigranular, and lithographic to very finely crystalline. They are generally somewhat siliceous and fossiliferous. Ostracod carapaces and sponge spicules are the predominant fossils, but fragmentary crinoids and unidentifiable fragments of fossils of other types are present. There is some pyrite in some of these limestones.

Near the Woodford contact, the Pinetop chert is more or less calcareous (dolomitic ?) and chalcedonic to cryptocrystalline (Pl. 4, fig. 5). Much of the chert is the "vesicular" type which is disseminated through the limy areas; it is brown in transmitted light and white in reflected light. This type of silica is characteristic of many limestone-replacement cherts. Colloform veins and nodules of chalcedony appear in places. Some nodules have centers of interlocked quartz grains and rims of chalcedony.

Woodford Chert (Devonian ?)

The Woodford chert unconformably overlies the Pinetop chert and is exposed in the block southeast of the Pine Mountain fault. According to Hendricks *et al.* (1947):

> "The basal 7 feet of the formation is a white chert breccia or conglomerate containing some lenses of very fine grained, crystalline limestone. The upper part consists of alternating beds of black papery shale that weather light gray, and black chert in beds 1 to 4 inches thick. It contains abundant round and discoidal phosphatic nodules and conodonts. The formation is about 67 feet thick."

Specimens from the lower zone consist of white calcareous spiculitic chert and chalcedonic chert. These rocks are intensely fractured and veined; some specimens are brecciated. Abundant euhedral to subhedral sublithographic to very finely crystalline carbonate rhombs are disseminated through the rocks. Much of the silica is in the isotropic state; it appears white in reflected light, brown in transmitted light, and dark under crossed nicols. The remaining silica is cryptocrystalline. The veinlets or fissure-fillings are both of quartz and chalcedony. There are a few large, nodular, colloform chalcedony bodies in the rocks. Sponge spicules are abundant, but other fossils are rare.

The shale in the upper part of the Woodford is dark red-brown and sapropelic. The shale is generally laminated and pyritiferous. By increase of silica it grades into sapropelic, spore-bearing chert. Most of the silica in the chert is apparently isotropic. Much of the chert is fractured and veined with fissure-fillings of quartz, chalcedony, and black asphaltic (?) material.

Embedded in the shale of the upper Woodford are a number of phosphate nodules. A thin section cut through a nodule shows irregularly shaped, replacement-type chert in a brownish, somewhat botryoidal, isotropic matrix of clay and phosphate. A few Radiolaria are present. The nodule contains phosphatic material of two shades of brown but there is little petrographic difference between the colored layers.

Chickachoc Chert (Pennsylvanian)

The Chickachoc chert lentil overlies the Spinger formation and is exposed in the block southeast of the Katy Club fault. According to Hendricks *et al.* (1947):

"The formation is generally poorly exposed with only the more resistant beds cropping out. Its thickness is estimated to be 600 feet. The formation consists of greenish gray to tan clay shale that weathers to a tan clay and contains beds and lenses of spiculite. The spiculite consists of a mass of sponge spicules cemented by siliceous limestone. Where it is fresh the spiculite is dark-bluish gray and very hard. Where it is weathered it resembles a coarse-grained sandstone or porous chert, but each grain is a sponge spicule. As many as ten beds of spiculite, each more than 5 feet thick, are present locally. In general, the base of a bed of spiculite, about 10 feet thick, is assumed to be the base of the formation; a bed of spiculite 20 to 50 feet thick is present in the middle part; and a very massive bed of spiculite, 30 to 120 feet thick, constitutes the top member. Locally, the Chickachoc chert contains abundant brachiopods, a few cephalopods, and other fossils of lower Pennsylvanian age that are identical with those of the Wapanucka limestone, with which it is correlated."

Thin sections of the Chickachoc consist largely of siliceous spiculite limestone or spiculite chert. Two major related lithologic types were observed. The first type is an inequigranular, very finely crystalline, fossiliferous, siliceous limestone. This rock contains abundant yellow chalcedony which is scattered through the rock in irregular masses. Glauconite and argillaceous material are present in a few specimens. The most abundant fossils are calcareous (and silicified) sponge spicules but there are some ostracods, crinoids, foraminifera, and unidentified fragmentary fossils.

The second lithologic type is a spiculitic chert consisting almost entirely of a felted mass of silicified sponge spicules (monaxons), with some interstitial clay, pyrite, and organic matter. It grades by increase of calcareous material into the first type, except that other fossils are lacking or rare.

Generally, in these rocks, the silica replaces the matrix, but in a few areas the calcareous sponge spicules are replaced selectively. All the sponge spicules appear to have been calcareous originally; upon silicification, brown chalcedony replaces the pore canals and clear chalcedony replaces the outside walls (Pl. 4, fig. 6).

Origin of the Siliceous Sediments

General Statement

One of the most controversial factors concerning the siliceous sediments of Ouachita facies is their origin. In particular, the source of the silica necessary for their formation has been a matter of argument. A considerable amount of geologic literature has been devoted to the origin of the Arkansas novaculite, with lesser emphasis placed on the other siliceous sediments.

Origin of The Bigfork Chert

Two major theories on the origin of the Bigfork chert are presented by Miser and Purdue (1929), and by Honess (1923). Miser and Purdue (1929, p. 129) state that:

"After the Womble deposition was completed, layers of silica and thinner layers of mud, aggregating 700 feet in thickness, were put down in probably all the Ouachita Mountain region and later formed chert and shale, respectively. To the formation thus produced the name Bigfork chert has been applied. . . . The water, to judge from the small amount of shale in the formation, was at first relatively clear, but late in Bigfork time much mud brought in from the land area was intermittently deposited with the silica forming the chert. Furthermore, the water was not of great depth, as is indicated by the ripple-marked surface of the sandstone. In the Bigfork, as in the Womble shale beneath, great numbers of graptolite remains were buried in the mud layers and are now well preserved in the shale. Evidently conditions at this time were not favorable for other kinds of marine life, as only a few brachiopods, bryozoans, and trilobites have been found in the Bigfork chert of Arkansas and a few sponge species in this formation in Oklahoma. The presence of the spicules suggests that at least a part of the silica which produced the chert was derived from such remains, but whether it was largely deposited by organic means or by chemical precipitation alone or by both processes is not known."

Honess (1923, pp. 79–80) considers that the Bigfork chert is a marine deposit, that limestone was more abundant originally in the Bigfork chert than now, and that much of the chert is secondary. He emphasizes the widespread distribution and uniform character of the Bigfork chert, the transitional contacts with the underlying and overlying formations, and the presence of crinoid stems and brachiopods in the cherty limestones.

It may be noted that Miser and Purdue favor a syngenetic or penecontemporaneous origin for most of the chert, whereas Honess prefers an epigenetic origin. However, the source of the silica must be considered first.

Widespread volcanism in the Ordovician period has been noted by C. S. Ross (1928), who states that:

> "Ordovician volcanic materials have been found by different geologists at many places in six of the eastern United States. . . . When all the evidence is considered, there can be no doubt that great volcanic eruptions occurred in Ordovician time and deposits hundreds of square miles apart show that thousands of square miles were at this time covered with ash."

We consider that ash falls were the source of silica for the Bigfork chert. The limestones in the lower unit of the Bigfork were apparently deposited in generally shallow, comparatively clear water with the clastic material derived from a very low-lying land mass (possibly predeposited and uplifted Ordovician shale and sandstone). Possibly 10–20 per cent of the silica in the lower limestones was syngenetic or penecontemporaneous. The remainder is considered to be epigenetic and in part a result of surface exposure. On the other hand, the dark, bedded cherty shales in the lower Bigfork were highly siliceous originally. They contain only radiolaria, spores, graptolites, and sponge spicules and were probably laid down in an euxinic environment unfavorable to bottom-dwelling organisms.

There is no sharp contact between the upper and lower units of the Bigfork and one thickens as the other thins. By upper Bigfork time, the seas were consistently muddy and the black shale environment obtained throughout. Volcanic ash continued to fall and was incorporated into the black chert and siliceous shales of the upper Bigfork. Probably not over 10–20 per cent of carbonate was ever present in the upper beds and this original carbonate has been replaced by silica in the outcrop occurrences. The Bigfork sediments were later subjected to the dynamic metamorphism of the Ouachita orogeny and some redistribution of the silica and most of the fracturing and vein-filling took place then.

Pettijohn (1949, p. 461–462) suggests that the differences between the Fernvale and Viola limestones of the Arbuckle Mountains and the equivalent Bigfork chert in the Ouachita Mountains result from silicification accelerated by deep burial and rise of geotherms and is not a facies change. We maintain that the petrographic data indicate that the major differences between the Fernvale-Viola and Bigfork formations are the result of facies changes. We consider the Fernvale and Viola sediments deposits of foreland facies and the Bigfork sediments deposits of geosynclinal facies.

The Viola formation of the Arbuckle Mountains contains considerable amounts of chert and siliceous limestone. Petrographic study strongly suggests that most of the silica epigenetically replaced limestone. In the Bigfork formation of the Ouachita Mountains, most of the silica is in the form of massive bedded chert and much of this silica is syngenetic or penecontemporaneous. Indeed, most of the upper part of the Bigfork chert consists of black, sapropelic spore-bearing cherts and black paper shales which were apparently never calcareous except in very limited degree. These black sapropelic graptolitic shales of the Upper Bigfork unquestionably are deposits of an euxinic environment and of themselves indicate a facies change from the limestones of the Viola and Fernvale formations.

Deep burial and consequent rise of geotherms must have occurred long after consolidation of the Bigfork formation (as the Ouachita geosyncline was not a major downwarp until Stanley-Jackfork time). Deep burial unquestionably modified the distribution of silica and may have caused some epigenetic silicification. However, the data indicate that most of the silica in the Bigfork formation is of primary origin and was derived largely from volcanic ash incorporated in the sediment at the time of deposition.

Origin of the Arkansas Novaculite

General statement.—The theories on the origin of the Arkansas novaculite have been summarized by Miser and Purdue (1929, p. 55–57). In chronological order, with one addition, these

theories are stated in generalized form in the following table[1]:

D. D. Owen (1868)—Metamorphosed sandstone.
J. C. Branner (1888)—metamorphosed chert.
T. B. Comstock (1888)—hot water deposits.
L. S. Griswold (1892)—sandstone of very fine grain.
Frank Rutley (1894)—siliceous replacement of dolomite or dolomitic limestone.
G. J. Hinde (1894)—organic origin.
O. A. Derby (1898)—replacement of limestone by silica.
W. H. Weed (1902)—chemical precipitate in deep sea.
C. R. Van Hise (1904)—organic precipitate, now recrystallized.
C. L. Baker (1918)—metamorphosed chert.
C. W. Honess (1923)—in part silicified and devitrified volcanic ash, but mostly chemical precipitation without the aid of organisms.
H. D. Miser and A. H. Purdue (1929)—chemically precipitated chert.
T. A. Hendricks *et al.* (1937)—an initially siliceous deposit derived at least in part from siliceous organisms.

Both Honess (1923, p. 138–139) and Miser and Purdue (1929, p. 56–57) give detailed reasons for the origin which they postulate and refute many of the other theories of origin. Part of their arguments are recapitulated in the following paragraphs.

Metamorphosed sandstone or very fine-grained sandstone theories.—There are occasional rounded "floated" quartz sand grains of detrital origin in the massive novaculite beds. Also there is every gradation from beds with an occasional sand grain to true sandstones which are both underlain and overlain by typical novaculite. Furthermore there are quartz grains which have been partially "resorbed" and incorporated into the novaculite so that their outlines are only faintly visible in plain transmitted light. Nevertheless, as Miser and Purdue (1929, p. 56) state:

"... The close-fitting arrangement of the grains of cryptocrystalline quartz in the highly calcareous and the chalcedony-like novaculite, which grade into each other, and the absence of any cementing material in these rocks indicate that the fine-grained quartz in them is not of detrital origin ... The absence of indications of any such pronounced regional metamorphism in the Ouachita region as would be necessary to have produced quartzite of a texture like the novaculite further supports this view. ..."

[1] See article by Miser and Purdue for original reference citations.

Miser (1943) discussed the regional metamorphism of the Arkansas novaculite and showed that, in areas of the Ouachita Mountains where the Arkansas novaculite is highly metamorphosed, as in the vicinity of Little Rock, Arkansas, the novaculite does not consist of dense cryptocrystalline silica, but the individual grains are increased in size and the rock resembles closely a metaquartzite.

Theory of organic origin.—This theory implies that the structureless cryptocrystalline silica which forms a large part of the Arkansas novaculite resulted from the death and solution of a great multitude of silica-bearing organisms, principally radiolaria and siliceous sponges. These organic remains are unquestionably very abundant in some beds of the Arkansas novaculite but form only a small percentage of the entire formation.

The essential fact which disproves this hypothesis is that most of the radiolarian tests and sponge spicules in the Arkansas novaculite are very well-preserved and show little indication of solution effects. Even in the metamorphosed novaculite of McCurtain County, Oklahoma, solution of the sheared and stretched-out radiolarian capsules has been quantitatively minor (Pl. 3, fig. 1). Furthermore, the dark novaculites at both Black Knob Ridge and in McCurtain County are generally sparse in siliceous organisms and all of their silica could not have resulted from solution of siliceous remains.

On the basis of available data, it must be concluded that the abundance of siliceous organisms in some beds of the Arkansas novaculite is a response to a favorable environment resulting from the high silica content of the waters in which the Arkansas novaculite was deposited.

Silicified limestone theory.—Proponents of this theory consider that the Arkansas novaculite was wholly, or to a large extent, limestone at the time of deposition. They believe that silicification was largely epigenetic, resulting from deep burial and intensification of metasomatic action in the Ouachita orogeny.

This theory may have been aided by the wide distribution of the whetstones made from the lower and calcareous upper members of the Arkansas novaculite. Petrographic study of slides made from this stone, without knowl-

edge of the field relations or lithology of the remainder of the formation, may have misled proponents of this theory.

The upper member of the novaculite in western Arkansas and in McCurtain County, Oklahoma contains many small rhombohedra and irregularly shaped aggregates of calcite and manganiferous carbonate. The carbonate is well-preserved only in unweathered specimens—where weathered out and removed by solution there is, at places, a residuum of hydrous iron oxides. Honess (1923, p. 130–134) considered that some of the opaque material replacing the carbonate is pyrolusite and that the original carbonate is rhodochrosite. He believed also that the carbonate was replacing the silica of the novaculite. Hendricks *et al.* (1937, p. 11) confirm this but state that:

"... the uniformly high silica content and the absence of any remnants of carbonate material other than clearly secondary crystals similar to those described by Honess leads to the conclusion that the novaculite is an initially siliceous deposit and not a replacement of limestone or other material."

There is yet another possible explanation for this lithologic type. Selectively silicified limestones are common at many places in the Paleozoic of the Mid-Continent region, and commonly silicification avoids the dolomite crystals in slightly dolomitized limestones. Minute dolomite rhombs may be perfectly preserved in rocks in which all the calcite has been replaced by chert. The carbonate rhombs in the upper member of the Arkansas novaculite may be relicts rather than a secondary replacement of the chert. It is not necessary to assume that the upper member of the Arkansas novaculite in McCurtain County and western Arkansas was deposited originally as limestone. However, it is considered that some carbonate was deposited syngenetically with the silica. Incipient dolomitization of the calcite may have occurred at places, followed by subsequent selective silicification of the calcite. This would account for many of the lithologic types not otherwise easily explained.

Honess's theory.—Honess states (1923, p. 138–139) that:

"... the lower and upper divisions of the Arkansas novaculite are essentially an accumulation of colloidal silica, precipitated on the sea floor, in some manner, through physical-chemical agencies, in marine waters which were more or less murky with fine black mud and silt most of the time but which were occasionally free from such matters ... The middle division of the novaculite formation ... is no different so far as the facts indicate from any other series of black cherts and black shales and need be explained in no other way."

Honess discussed the "volcanic ash" in the lower novaculite at Cypress Creek, McCurtain County, Okla., but ruled out extension of this idea to the formation as a whole because of the uniformly high silica content of most of the novaculite.

Theory of Miser and Purdue.—Miser and Purdue (1929, p. 57) consider that the novaculite is a chert, and state that:

"... the silica which formed this quartz was deposited in the sea simultaneously with the other material of the formation ..., because the novaculite is overlain, underlain, and in part interbedded with beds of clay shale so impervious that they would have prevented the free circulation of the underground water necessary to introduce from other sources the large amount of silica required as replacement material. To be sure, the extensive jointing of the shales has made them somewhat pervious, so that in places the underground water circulates in them freely, but the usual even distribution of the silica everywhere in any particular layer or bed of the novaculite precludes the possibility that it was introduced subsequent to the jointing. Besides, the jointing of the novaculite beds themselves shows that these beds were siliceous when the jointing took place."

They conclude that the source of the silica was chemical precipitation.

Theory promulgated in this paper.—The lack of agreement as to the origin of the Arkansas novaculite probably resulted from insufficient petrographic study of the formation as a whole. Actually, the Arkansas novaculite is a highly variable complex of sediments and many factors were involved in its origin. The most important and controversial factor is the source of the silica. We consider that Devonian-Mississippian volcanism resulted in extensive ash falls during a time of relative peneplanation when little detrital sediment was being supplied to the Ouachita geosyncline. Submarine weathering (halmyrolysis), prolonged over long periods of time, removed readily soluble elements and converted the ash to nearly pure opaline or "isotropic" silica. Radiolaria and siliceous sponges thrived in the high-silica environment and their remains are preserved in the siliceous sediments. There was probably a normal marine

environment during most of the time of deposition, but occasionally stagnant bottom conditions prevailed and sapropelic high-silica sediments containing spore exines were deposited. Intermittent increases of stream gradient caused the deposition of clastic detritus far out in the geosyncline and resulted in the interbedded shales and scattered thin sandstones. Some parts of the upper Arkansas novaculite must nave been deposited in subaerial or possibly tidal flat conditions, as stumps of *Callixylon* standing in an upright position have been observed at several places. Furthermore one set of thin sections of a chert nodule from the upper part of the novaculite in the Potato Hills reveals sheaflike crystal aggregates that were considered by Waldemar Schaller (personal communication) to be silica replacements of an undetermined saline mineral that grew in a mud. Such material might be expected to form on a tidal flat.

Colloidal isotropic silica, under the influence of deep burial for long periods of time, would tend to lose water and crystallize into chalcedonic and cryptocrystalline silica. This tendency was reinforced by geosynclinal downwarping and the subsequent Ouachita orogeny. It is considered that the novaculite of Black Knob Ridge is only slightly metamorphosed and closely approximates the lithology of the original sedimentary sequence, whereas, as shown by Miser (1943), the novaculite of McCurtain County and western Arkansas has been subjected to strong dynamic metamorphism.

Origin of the Siliceous Shales

It is considered that the silica of the siliceous shales of the Stanley, Jackfork, and Atoka formations was derived from submarine weathering of volcanic ash, and that they were probably deposited in a lime-poor habitat. The restricted fauna of siliceous organisms (radiolaria, sponge spicules) and chitinous and phosphatic organisms (linguloid and orbiculoid brachiopods, conodonts) suggest that the sea in which the siliceous shales were laid down was high in silica and unfavorable to calcareous organisms. Many of the siliceous shales are thinly varved, suggesting slow deposition in quiet water. There are rapid alternations at some places from dark, carbonaceous or sapropelic, pyritiferous siliceous shale deposited in a reducing environment to glauconitic siltstone containing fossils that were originally calcareous and apparently deposited in a normal marine environment. Large amounts of finely macerated to fairly large fragments of plant material are present locally in the siliceous shales and are common in the associated sandstones and shales. Sandstone dikes have been noted in the Moyers siliceous shale.

The lenses of chert conglomerate intercalated at places in the Moyers, Chickasaw Creek, and Wesley siliceous shale may be highly significant. These intraformational conglomerate lenses include a variety of rock fragments, but granules and sand-size fragments of siliceous shale and chert similar in lithology to the enclosing sediments predominate. This suggests that deposition was just below wave base and that rare exceptional storms (or minor changes in sea level) broke up and eroded the shales and incorporated the resultant fragmentary particles into the later shale layers.

In addition to intermittent eruptions of volcanic ash, a brief hiatus in the supply of coarse detrital sediment may have been involved in the formation of the siliceous shales. They are much finer-grained than the surrounding sediments, although they may grade abruptly into sandstone or siltstone. This suggests that there were short periods of quiescence in the early stages of the sinking and deformation of the Ouachita geosyncline, allowing the ash falls to be concentrated within a few feet of slowly deposited strata rather than disseminated over hundreds of feet of more quickly deposited coarser sediments.

Summary

Although there are conspicuous lithologic differences between the Bigfork chert, Arkansas novaculite, and the siliceous shales of the Stanley, Jackfork, and Atoka formations, certain characteristics suggest a common or related origin. Some of these are:

1. Calcareous fossils are rare whereas siliceous and chitinous forms such as conodonts,

graptolites, sponge spicules, and radiolaria are very abundant in some beds.

2. Normal sedimentary deposits such as sandstones, limestones, and gray marine shale are subordinate in the siliceous formations to unusual rock types such as radiolarian chert, spiculitic chert, novaculite, sapropelic chert, sapropelic black shale, and siliceous shale.

3. All of the siliceous sediments appear suddenly, generally overlying normal clastic sediments of geosynclinal facies, and pass gradationally in some cases into the beds overlying them.

4. Field evidence suggests that all the siliceous sediments were deposited in relatively shallow water which became stagnant at times.

5. All the siliceous sediments maintain their general lithologic character over long distances.

It is considered that the siliceous sediments of Ouachita facies formed as a result of extrusive vulcanism and that the source of their silica was submarine weathering (halmyrolysis) of volcanic ash. All of the siliceous sediments apparently were deposited at times when relatively little clastic sediment was being supplied to the "Ouachita geosyncline." The Hatton tuff lentil in the lower part of the Stanley shale is a deposit of volcanic ash possibly similar to the parent material of the cherts and novaculites, but it was laid down during a time of active downwarping, was buried quickly by clastic sediments, and was never converted to chert. The abundance of siliceous organisms and the lack of calcareous organisms in the siliceous sediments resulted from the high-silica environment. The abrupt lower contacts and gradational upper contacts of the siliceous sediments resulted from a sudden beginning and spasmodic diminution of volcanic activity. The dominance of unusual sedimentary types in these formations is due in part to the abundant supply of volcanic ash, and in part to the fact that little clastic material was being derived from the low-lying land-masses of Bigfork and Arkansas novaculite times. The siliceous shales of the Stanley, Jackfork, and Atoka formations apparently represent minor recurrences of waning Devonian-Mississippian vulcanism, each possibly accompanied by a brief hiatus in the deposition of coarse clastic sediments.

Subsequent to their deposition, all the siliceous sediments have been affected by epigenetic redistribution of part of their silica, partly because of the dynamic metamorphism of the Ouachita orogeny, and partly because of surface exposure.

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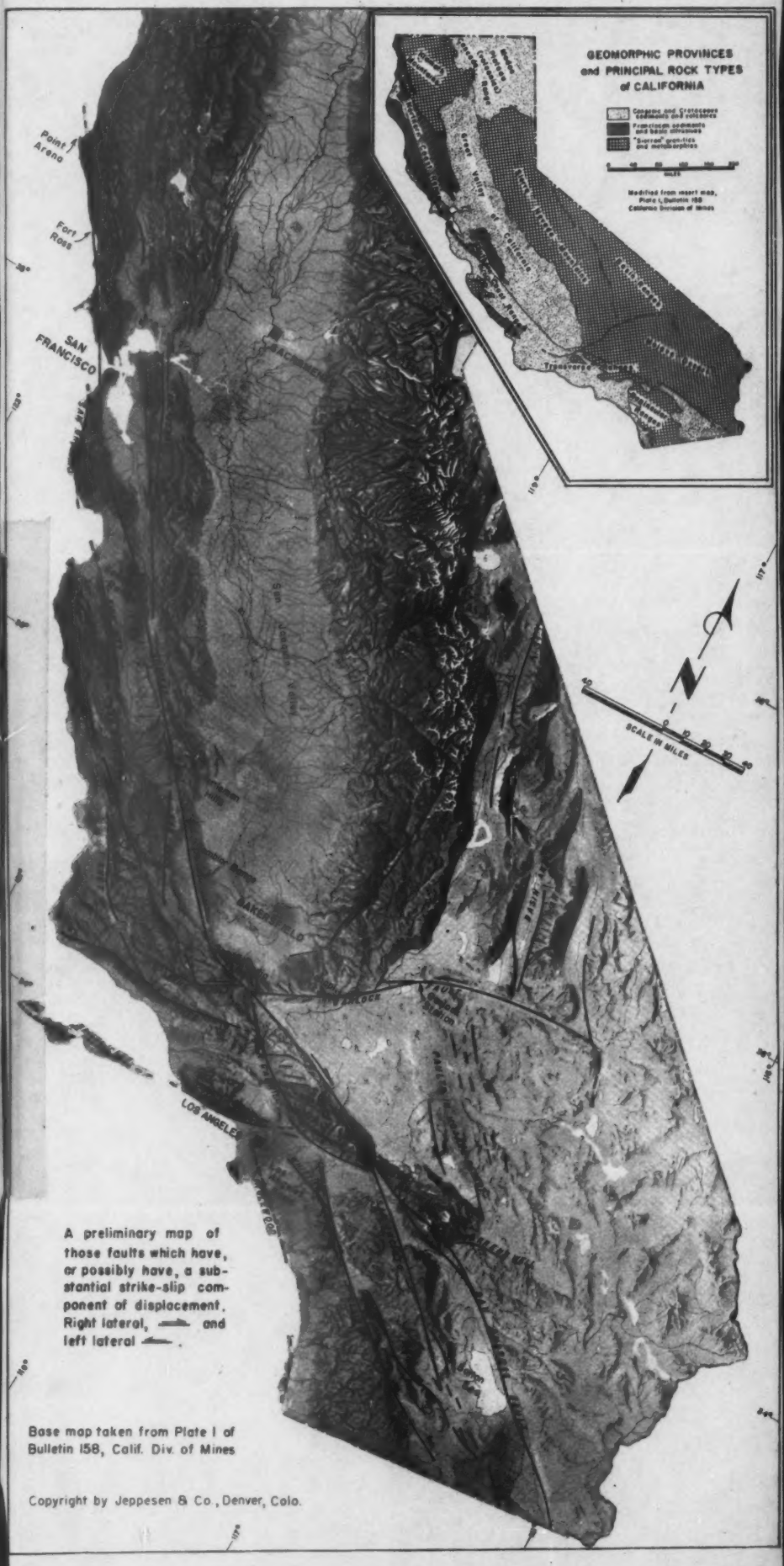
GEOMORPHIC PROVINCES
and PRINCIPAL ROCK TYPES
of CALIFORNIA
Modified from insert map, Plate I, Bulletin 158 California Division of Mines
Point Arena
Fort Ross
SAN FRANCISCO
SACRAMENTO
BAKERSFIELD
LOS ANGELES
SCALE IN MILES
A preliminary map of those faults which have, or possibly have, a substantial strike-slip component of displacement. Right lateral, and left lateral.
Base map taken from Plate I of Bulletin 158, Calif. Div. of Mines
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LATERAL FAULT MAP OF CALIFORNIA

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SAN ANDREAS, GARLOCK, AND BIG PINE FAULTS, CALIFORNIA

A Study of the Character, History, and Tectonic Significance of their Displacements

By Mason L. Hill and T. W. Dibblee, Jr.

Abstract

The Big Pine left lateral fault extends northeastward from Big Pine Mountain to the right lateral San Andreas fault, while the left lateral Garlock fault extends northeast from the San Andreas, but from a point 5 miles to the southeast. The Big Pine fault is considered the western segment of the Garlock fault as offset by the San Andreas. Movement on this Garlock-Big Pine fault zone appears to have caused the anomalous east-west trend of the San Andreas fault in this vicinity.

Tens of miles of lateral movement have probably occurred on these faults with the possibility of a cumulative movement on the San Andreas of hundreds of miles since Jurassic time. Such distances are important elements in reconstructing paleogeologic conditions.

The three concurrently active, long, steep, and deep faults are considered major conjugate shears which define a primary strain pattern of relative east-west extension and north-south shortening of an area of approximately 120,000 square miles. A northeast-southwest counterclockwise compressive couple, possibly set up by drag due to the deep-seated movement of rock material from the Pacific region, is tentatively postulated as causing the deformation in this large region.

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Introduction

Recent work shows, contrary to previously published maps, that the Big Pine fault, on the south slope of Big Pine Mountain, extends northeastward to the San Andreas. This work was done by the junior author (Dibblee) in 1947, on photographic and topographic base maps, as part of an oil-exploration program. The characteristics of the fault were recognized as possibly of regional tectonic significance which stimulated further studies by Dibblee and Hill (1948) upon which this presentation is founded.

The first part of the discussion presents evi-

dence on the character and amount of the displacements of the faults with particular emphasis on interpretations of the history of the San Andreas fault; the second part deals with interpretations of relationships between these faults and their bearing on some other structural features of the region; and the third part analyzes the mechanics of the deformation and timidly suggests some genetic concepts and conclusions.

Most of the facts concerning the faults discussed are generally known, but some new data and interpretations lead to conclusions, particularly concerning the San Andreas fault, that are not now orthodox. These conclusions are mainly tentative and are offered for the criticism of other workers.

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San Andreas Fault

Problems

This great Right Lateral[1] fault became widely publicized after the San Francisco earthquake of 1906. At that time a maximum of 21 feet of right lateral offset occurred on the San Andreas fault (Lawson, 1908). Much geologic study has since been devoted to this fault, and it has been mapped from Point Arena southeastward almost to Mexico, a distance of about 600 miles; however, the history of its displacements is still controversial. After nearly 50 years of research, workers disagree in their answers to the following questions:

1. What is the San Andreas fault?

Noble (1933, p. 11 says:

"It is marked by a curiously straight and almost continuously traceable chain of scarps, ridges and trough-like depressions. . . . This line of recent topographic features upon the San Andreas fault is commonly referred to as the fault trace. Bordering the fault is a belt of roughly parallel branching and interlacing fractures which at some places attains a width of several miles. This belt, the San Andreas fault zone, is a mosaic of elongated sliver-like blocks or wedges whose longer axes trend parallel with the main fault, so that the dominant structure is a sort of slicing. . . ."

On the other hand, Taliaferro (1943, p. 160) says:

"Nearly everywhere along the San Andreas there is abundant physiographic evidence of recent faulting, such as true sag ponds and offset ridges and drainage lines. However, along the earlier mid-Pleistocene faults (which the San Andreas partially follows) there is no similar direct physiographic evidence of faulting. . . . It (the San Andreas) has produced no important modifications of either the structures or topography formed by these (earlier) diastrophisms. . . . The supposed branches or 'barbs' are actually earlier faults which were formed by a very different type of movement and which may be traced across the San Andreas."

2. What is the age of the San Andreas fault?

All workers agree, on the firm basis of earthquake activity and topographic expression, that this fault is active. However, there is a distinct difference of opinion, in respect to its antiquity.

Noble (1933, p. 11) says:

"The fault is a very old line of weakness, upon which movements have recurred through Tertiary and Quaternary time and perhaps through much of pre-Tertiary time."

Taliaferro (1943, p. 161) contends that the San Andreas fault had its inception in late Pleistocene time and that it, with its unique characteristics, should not be confused with the older faults which are followed, and occasionally crossed, by the San Andreas.

3. What is known about the orientation (sense) of displacements of the San Andreas fault? All authors have agreed that the latest movements have been actually strike-slip with the southwest block relatively moving to the northwest (right lateral), but here again agreement stops. Those workers who believe that the San Andreas is an old fault differ in opinion on the relative importance of vertical and lateral components of these earlier movements. Most authors, for example Buwalda (1926), Clark (1929), Weaver (1949), and Willis (1929), believe that these earlier movements were predominantly vertical and that reversals of throw have taken place on this and parallel faults

[1] The block opposite the observer, as viewed in horizontal section, is offset to the right (Hill, 1947). The terms *right* and *left lateral* are as critical in fault nomenclature as *normal* and *reverse*; they are descriptive of *separation* rather than definitive of *slip*; they are geometric rather than generic; and they are becoming widely used. Some English workers (Anderson, 1942, p. 55) use *dextral* and *sinistral transcurrent* for these types but with a *slip* instead of *separation* connotation.

4. Does the San Andreas fault zone mark a prominent contact between unlike rocks?

Noble (1933, p. 11–12) says:

"The profound difference in the rocks on opposite sides of the San Andreas fault shows that the fault movements have been of great magnitude. . . . Nowhere in the 30 mile sector (north side of San Gabriel Mountains) are the rocks on opposite sides of the fault similar."

Taliaferro (1943, p. 160) says:

"The San Andreas fault, however, is rarely the boundary between these two very diverse types, but usually is either wholly within crystalline (granitic) rocks or Mesozoic (Franciscan) rocks, except where it cuts through either Miocene, Pliocene or Pleistocene sediments."

5. What is the order of magnitude of the cumulative offset on the San Andreas fault?

Noble (1925–1926, p. 420) says early Tertiary strata may have been offset by the San Andreas (in a right lateral sense), a total distance of approximately 24 miles.

Taliaferro (1943, p. 161) says:

". . . . The horizontal shift (on the San Andreas) has been small, and has not been greater than one mile and probably even less."

The present authors believe that: (1) the San Andreas is a steep fault zone of variable width consisting of one or several nearly parallel faults; (2) its inception was quite likely pre-Tertiary, and it is now active; (3) it has probably been characterized by right lateral displacements throughout its history; (4) it marks such an important contact that rarely can it be crossed, except in Recent alluvium, without passing into significantly different rocks; and (5) its cumulative displacement of some rock units is at least tens of miles, and older rocks may have been displaced a few hundred miles.

Description and Interpretations

The San Andreas fault is so long that no one worker can have a first-hand detailed knowledge of its characteristics throughout its length. The junior author, however, has mapped a 100-mile stretch of the fault and associated geology in the Salton Sea region and most of the 300-mile stretch from the San Emigdio Mountains to San Francisco. This work and available maps and discussions by others are used in the following outline of significant fault characteristics and their bearing on the history of the San Andreas fault.

1. The trace of the San Andreas zone is typically continuous and straight (Pl. 1). Topographic and geologic features clearly show its continuity for about 600 miles. There is evidence of recent activity along its entire course. Excepting a 30-mile segment trending eastward in the San Emigdio Mountains, and another stretch of similar trend 100 miles to the southeast, the zone is remarkably straight from Point Arena southeastward nearly to Mexico. These aspects of continuity and straightness are considered typical of strike-slip faults.

2. The San Andreas is a steep fault which transects major topographic features but develops all along its course one or several parallel trenches, sag ponds, low ridges, saddles, and/or scarps. Its steepness is indicated by the straight trace, the fact that mapped fault planes are nearly vertical, and the failure of near-by drill holes to penetrate the zone. These characteristics are typical of strike-slip faults. The development of fresh topographic features, many of which are in unconsolidated recent sediments, and the common lack of appreciable vertical or consistent vertical components of offset clearly indicate the recency of lateral movements. Seismic evidence for recent right lateral movements on the San Andreas, as summarized by Wallace (1949), comprises the following maximum displacements at the time of earthquakes: 30 feet (San Emigdio Mountains, 1857), 10 feet (San Francisco area, 1868), 21 feet (San Francisco area, 1906), and 10 feet (Salton Sea area, 1940). Geodetic surveys in the San Francisco (Whitten, 1948) and Salton Sea (Meade, 1948) areas also show this type of recent movement.

3. The San Andreas fault zone ranges from a few feet to a few miles in width. Locally a single recent trace may be irregular, with 15-degree variations in strike within a few hundred feet, or it may disappear and be replaced, en echelon, by another. Occasionally two or three parallel traces widen the zone of recent traces to a maximum of about half a mile. Wider segments of the zone consist of several faults (not necessarily active) which are usually steep and nearly parallel to the trend of the zone. These

characteristics are considered typical of strike-slip fault zones along which recurring movements have taken place.

4. The apparent throw is commonly reversed along the San Andreas fault as indicated by topographic and geologic relationships. These throws are probably due to the major strike-slip component which places in juxtaposition unlike topographic elevations and geologic sections, and thus the reversals of dip-slip are mainly illusory.

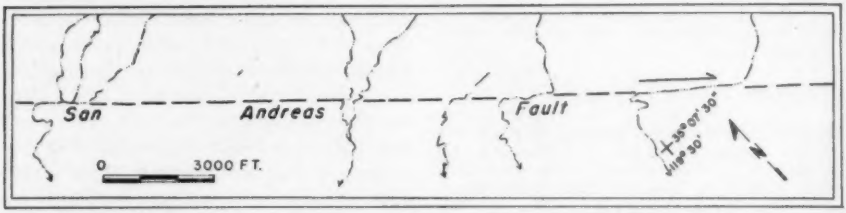

FIGURE 1.—RIGHT LATERAL OFFSET OF DRAINAGE LINES BY THE SAN ANDREAS FAULT
Elkhorn Hills quadrangle

5. Drainage lines are consistently offset in a right lateral sense. These offsets are especially clear on the southwest side of the Temblor Range (Fig. 1) where a maximum of 3000 feet of displacement has occurred through recent movements on the fault. Wallace (1949, p. 805) reports a probable drainage offset of 1½ miles on the north side of the San Gabriel Mountains, and Allen (1946, p. 50) reports 3800-foot offsets of drainage lines near the Gabilan Range, also in a right lateral sense.

6. Recently developed trenches which trend southward into the fault have been observed in aerial reconnaissance on the southwest side of the Temblor Range. These are oriented correctly to be tensional in origin and due to right lateral movement on the San Andreas.

7. Locally developed west-northwest trending folds adjacent to the San Andreas are obviously drag folds resulting from the right lateral movement on the San Andreas. Such drag folds are especially clear in the Salton Sea Region, and, besides indicating the right lateral sense of movement on the fault, many of them show by their discordance with topographic form that the fault was active before the present physiographic features were developed.

8. Wallace (1949) reports a probable 6-mile right lateral offset of terrace deposits on the north side of the San Gabriel Mountains, and L. F. Noble (personal communication) describes similar late offsets in that area of several miles.

9. Between the San Emigdio Mountains and the Temblor Range, there are two facies of Pleistocene gravels. On the southwest side of the San Andreas, the pebbles are granite, gneiss, quartzite, limestone, black shale, and sandstone which undoubtedly came from the San Emigdio Mountains. On the other side of the fault, the pebbles are almost exclusively white siliceous shale which probably came from the Miocene shale of the Temblor Range. These two facies are in direct contact along the San Andreas for several miles. Furthermore, the northwest end of the crystalline clast facies is about 14 miles northwest of the crystalline rocks of the San Emigdio Mountains. These relationships thus indicate a right lateral displacement of approximately 10 miles on the San Andreas fault since Pleistocene deposition in this area (Fig. 2).

10. In the Caliente Range, marine sediments of upper and middle Miocene age grade laterally eastward into continental red beds which strike into the San Andreas fault, whereas strata of the same age are marine shales on the other side of the fault. This juxtaposition of unlike facies again demonstrates substantial lateral movement. In this case the general trend of the western margin of the continental facies in the Caliente Range is northward across the Carrizo Plain toward the San Andreas, whereas possibly the same transition line may be extrapolated southward from along the east side of the San Joaquin Valley to the fault. Thus, by simple projections the right lateral offset on the

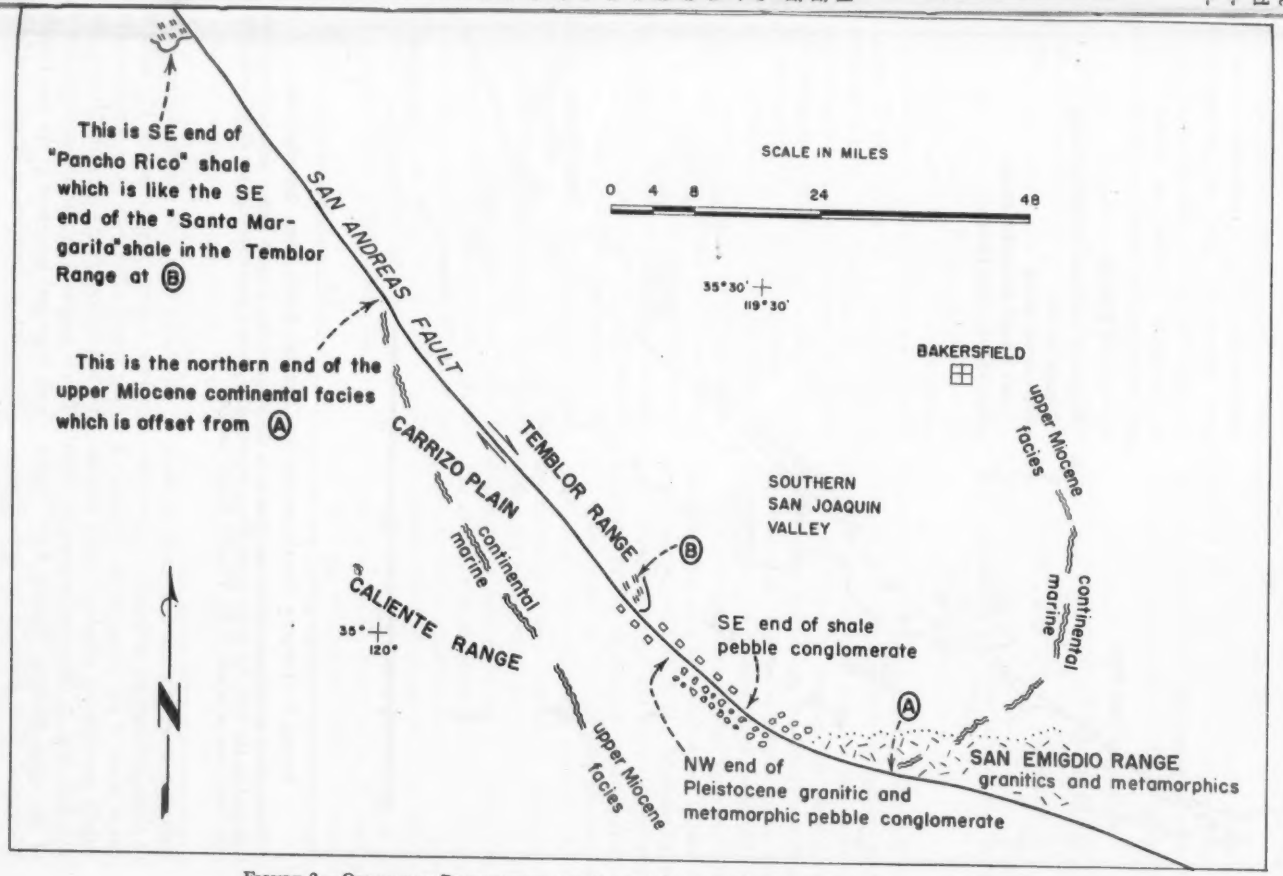
This is SE end of "Pancho Rico" shale which is like the SE end of the "Santa Margarita" shale in the Temblor Range at B
This is the northern end of the upper Miocene continental facies which is offset from A
SAN ANDREAS FAULT
SCALE IN MILES
0 4 8 24 48
35°30' 119°30'
BAKERSHELD
upper Miocene facies
continental marine
SOUTHERN SAN JOAQUIN VALLEY
CARRIZO PLAIN
TEMBLOR RANGE
B
continental marine
CALIENTE RANGE
35° 120°
SE end of shale pebble conglomerate
A
SAN EMIGDIO RANGE
granitics and metamorphics
NW end of Pleistocene granitic and metamorphic pebble conglomerate
upper Miocene facies
N

FIGURE 2.—OFFSETS OF PLEISTOCENE AND UPPER MIOCENE FACIES BY THE SAN ANDREAS FAULT

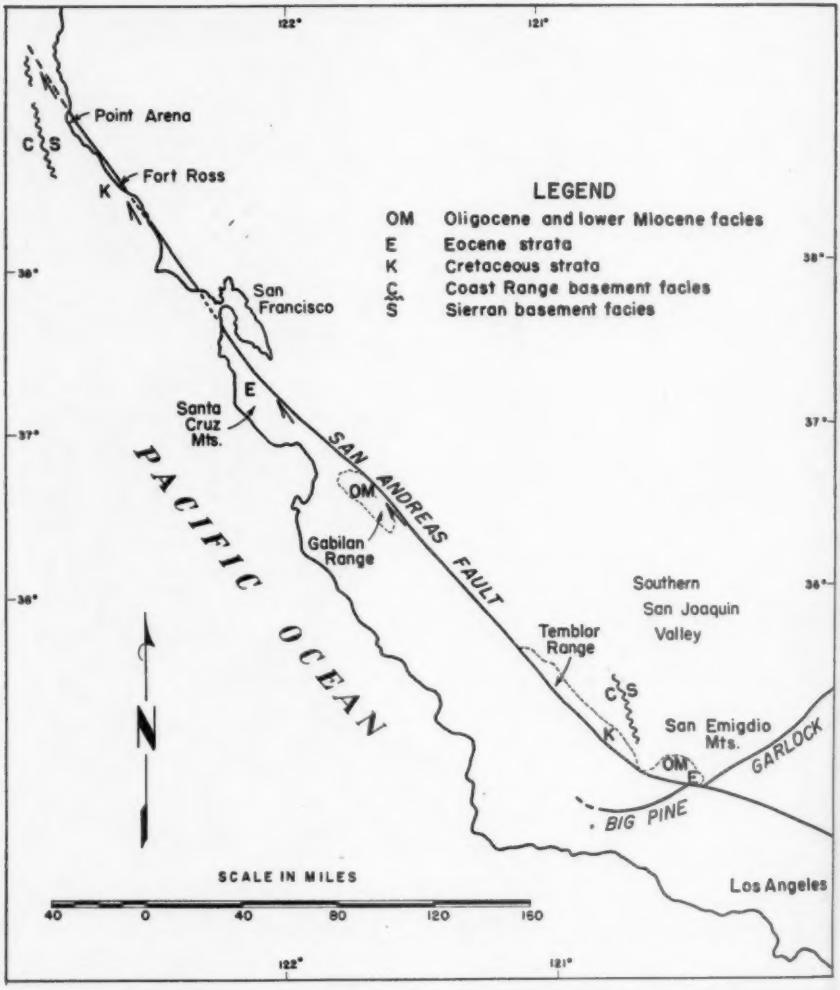

FIGURE 3.—POSSIBLE OFFSETS ON THE SAN ANDREAS FAULT SINCE JURASSIC TIME

These time-rock facies appear to be offset from the southern San Joaquin Valley area to the positions indicated on the west side of the fault. Thus, for example, during Cretaceous time the Fort Ross section may have been adjacent to the Temblor Range section.

fault since upper Miocene time would be about 65 miles, although the probability of irregularities in trend of this facies contact precludes a strictly quantitative solution of that cumulative shift (Fig. 2). Note the comparable offset of the upper Miocene "Pancho Rico"-"Santa Margarita" shale, shown in the same figure.

11. Going back only slightly farther in the geologic record, approximately 175 miles of right lateral offset may have accumulated on the San Andreas fault since early Miocene time. This is suggested by the unique similarities of rock types and sequences in the San Emigdio Mountains, as described by Wagner and Schil-

ling (1923), and the Gabilan Range as described by Kerr and Schenck (1925), and Allen (1946). In each of these areas, a section of lower Miocene volcanics, red beds, and marine lower Miocene and Oligocene strata occurs (Fig. 3).

12. A similar relationship is suggested by some lithologic and faunal similarities between the Eocene formations of the Temblor-San Emigdio and the Santa Cruz Mountains which indicate the possibility of an offset of approximately 225 miles since late Eocene time (Fig. 3).

13. Also the southern limit of Cretaceous strata in the Temblor Range may match with the southern limit of Cretaceous beds near Fort Ross which would indicate an offset of approximately 320 miles (Fig. 3). This possibility is strengthened by the improbability that vertical movements were consistent enough to remove the Cretaceous strata from such a long belt on the southwest side of the San Andreas fault zone. Furthermore, the Cretaceous sediments which do occur as close as 12 miles southwest of the San Andreas in the La Panza Range (Pl. 1) are west of other important faults and rest on granitic instead of Franciscan basement.

14. Of great interest would be good evidence for offset of the pre-Cretaceous rocks by the San Andreas. Perhaps the southward-trending contact between the Sierran complex and the Coast Range Franciscan complex which is concealed by the sediments of the Great Valley is truncated by the fault in the southern Temblor Range with its offset equivalent concealed by the sea north of Point Arena. This would amount to at least 350 miles of right lateral displacement since Jurassic time (Fig. 3).

Discussion and Conclusions

The antiquity and amount of movement on the San Andreas fault is evidenced by the increased offset of older rock facies as matched across the fault zone. The possibility of a pre-Cretaceous age and a cumulative right lateral displacement of several hundred miles is thereby suggested. These conclusions are both speculative and qualitative but tenable enough to justify further critical work to test their correctness and to determine more exactly their values. There is, however, no reasonable doubt that the fault is at least as old as early Pliocene and that the total right lateral offset is at least tens of miles (Figs. 2, 3).

Of perhaps great significance is the increased offset of older rock facies on the west side of the San Andreas from their probable counterparts at the southern end of the San Joaquin

FIGURE 4.—SKETCH SECTION ACROSS THE BEAR VALLEY FAULT

After Wilson (1943, p. 227)

Valley (Fig. 3). This situation appears to indicate that one block was actually moving past the other. If this were not true the counterparts would also be progressively shifted in the opposite direction. A conclusion regarding this movement will be presented under the subject heading of "Mechanics of the faulting."

Those workers who maintain that the San Andreas refers only to a fault which has been active recently and which had its inception in late Pleistocene time and those others who believe in earlier dip-slip and reversals of dip-slip displacement on the San Andreas may be skeptical of these conclusions. However, these differences in opinion appear to result from definitions of the fault and interpretations of displacement from cross-section relationships. The present authors believe that: (1) in many places the San Andreas is a wide zone comprising several important faults which are parallel or nearly parallel to the recent trace, and (2) the apparent dip-slip displacements are mainly the result of substantial lateral movements which put unlike geologic sections in juxtaposition.

The Bear Valley fault, east of the Gabilan Range (Wilson, 1943, p. 251), illustrates these differences in opinion and interpretation. This fault is mapped about half a mile southwest of the recent trace of the San Andreas and dips steeply northeast. It separates middle Miocene (Monterey) sediments resting on the Santa Lucia granite on the southwest side from a section of Plio-Pleistocene (San Benito) gravels resting on the Franciscan complex (Fig. 4). Wilson's interpretation is that this "ancestral"

San Andreas fault is not the San Andreas because it shows no evidence of recent movement, it is not vertical, and it is supposed to be characterized by dip-slip rather than strike-slip displacement. According to Wilson the first movement, probably Eocene normal faulting, elevated the west side causing the Franciscan to be removed by erosion. The second period of movement was post-Mjocene, and, by reverse faulting, the east side was elevated with accompanying erosion of all the Miocene strata on that block. The third period of movement occurred in late Pleistocene or Recent times as normal faulting which again elevated the west block causing the removal of the San Benito gravels.

The present authors suggest, based on the same data and by pinning a little faith on the law of uniformitarianism, the following tentative interpretations:

The juxtaposition of granitic and Franciscan basement may be the result of a great many miles of cumulative lateral movement on the fault. The absence of Miocene sediments on the northeast side of the fault may be due to many miles of right lateral movement so that the facies of the southwest block in this area were offset far to the southeast. The absence of the Plio-Pleistocene gravels on the southwest side of the fault in this area may also be due to right lateral offset with the possibility that they could be found several miles to the northwest on that side of the San Andreas. Although no recent movement has occurred on the Bear Valley fault, it is probably an inactive and slightly deformed fracture within the San Andreas zone which was characterized by the same type of movement as is evidenced along the recent trace of that fault.

It appears then that the significance of the Bear Valley fault is debatable. It is discussed here only as an example. An alternate interpretation is suggested in order to focus attention on a possibility that seems to have been overlooked. If the Bear Valley fault or other structures similarly related to the San Andreas, such as the Pilarcitos fault in the San Francisco area, can be shown to effect substantial right lateral offset, it will be reasonable to include them in the San Andreas fault zone as a partial manifestation of the important history of that zone. Furthermore, it would eliminate the not too probable interpretations of reversals of dip-slip displacements from cross-section relationships and the postulation of a unique and new strain-stress environment to explain the recent trace of the San Andreas fault.

The type of evidence used in this paper for the cumulative offset on lateral faults is not wholly new, nor are the suggested offset distances unprecedented. Kennedy (1946) has demonstrated 65 miles of strike-slip displacement on the Great Glen fault in Scotland, and the Alpine fault in New Zealand may have 300 miles of lateral displacement according to Wellman (1949 and personal communication). Even in California Vickery (1925) determined that the relatively minor Sunol (Calaveras) fault has had 13 miles of right lateral displacement since early Miocene time by what appears to be a unique match of strata across this fault. Furthermore, the cumulative amount of lateral movements, which are tangential to the earth's surface, need not be as limited as for high-angle dip-slip faults which are necessarily limited by gravitational forces.

Much more work is obviously needed to verify qualitatively some of these tentative conclusions. Many more possible offsets must be analyzed for accuracy and compatibility with respect to geologic time and geographic distribution before sound quantitative conclusions can be reached. In general, the younger rocks should be more easily matched across the fault, but the offsets of older units, particularly the pre-Cretaceous crystalline rocks, will be more significant. For example, is the north end of the Pelona schist on the north side of the San Gabriel Mountains (Wallace, 1949) offset 160 miles from the north end of the Orocopia schist near the Salton Sea (Miller, 1944)? Then, is such an offset commensurate with the pos-

PLATE 2. SAN ANDREAS, GARLOCK, AND BIG PINE FAULTS

FIGURE 1. TERMINATION OF BIG PINE FAULT AT THE SAN ANDREAS
View north

FIGURE 2. TRENCHES ADJACENT TO GARLOCK FAULT

FIGURE 1

FIGURE 2

SAN ANDREAS, GARLOCK, AND BIG PINE FAULTS

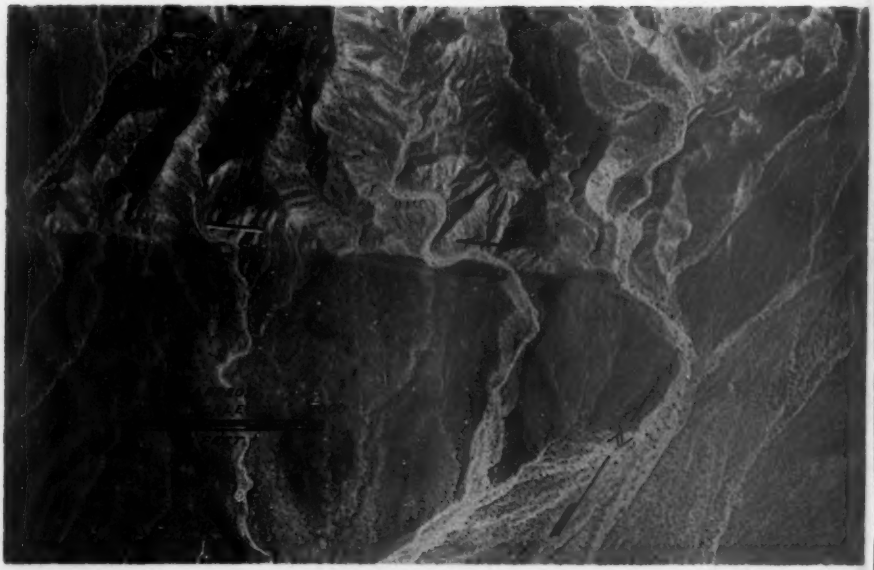

FIGURE 1

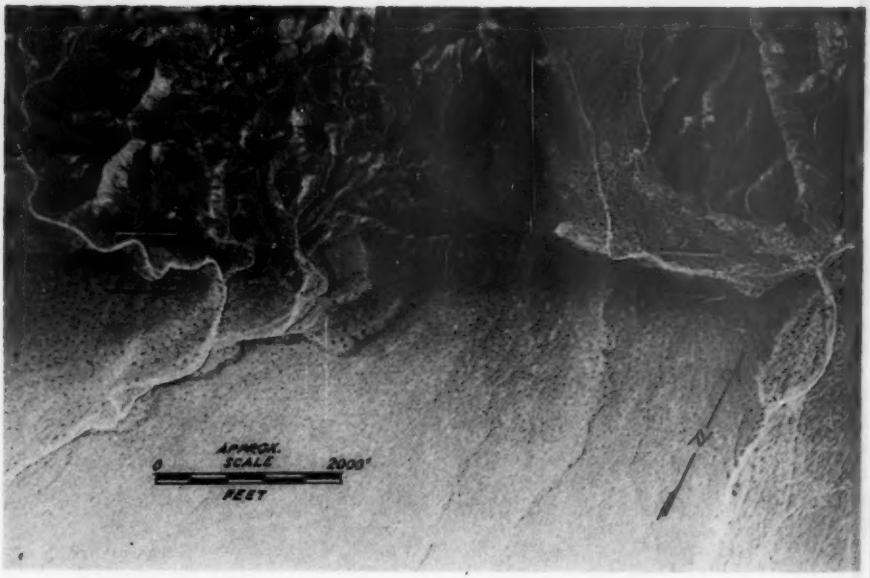

FIGURE 2

GARLOCK FAULT

sible offset on the San Andreas north of the Garlock-Big Pine fault zone?

Another worth-while line of research is the study of rates of movement on the San Andreas fault. Geodetic work has demonstrated a right lateral movement in the San Francisco Bay region averaging 2 inches a year since the survey of 1880 (Whitten, 1948). Another method, based on offsets at the time of earthquakes, suggests an average movement along the entire length of the fault of about 0.2 inch a year since 1857 (Wallace, 1949). Now, from the probable and possible offsets described above and the available calculations of geologic time in years, the rates of movement come out in fractions of an inch per year, as follows: 0.25 since Miocene, 0.2 since lower Miocene, 0.3 since Eocene, 0.29 since Cretaceous, and 0.2+ since Jurassic time. These figures are surprisingly consistent and therefore may possibly be taken as evidence that the rate of movement has been rather uniform for some 100 million years. On the other hand, the geodetically determined rate is nearly tenfold greater than these geologically controlled figures. However this great rate might be anomalous because of the short time involved, the short length of the fault involved, the inclusion of the large displacement of 1906, and/or movements outside the fault zone.

Garlock Fault

The Garlock fault extends from the San Andreas northeast and east-northeast for 150 miles (Pl. 1). It, like the San Andreas fault, transects topographic features, dips steeply, and has apparent variations of throw along its course.

Hess (1910, p. 25) first described and named the fault and indicated its type locality as just north of Garlock Station. Hulin (1925) was the first to suggest lateral displacement. He used the offset of a contact as possible evidence for left lateral displacement of approximately 6 miles.

The present authors followed this fault zone from its junction with the San Andreas northeastward for 150 miles by airplane and saw many well-defined offsets of drainage lines. These are all in a left lateral sense and commonly a distance of 2000 feet (Pl. 3). Also seen on this air reconnaissance were trenches in old alluvium in the vicinity of Garlock Station, immediately north of the fault trace (Pl. 2, fig. 2). These parallel trenches, approximately 40 feet deep, 150 feet wide, and 2000 feet long, trend S. 30°W. into the N. 60°E. trending Garlock fault trace. Their characteristics, including orientation and proximity to the Garlock fault, suggest gash fractures produced by tensional stresses adjacent to a major shear zone.

These striking evidences of recent and actual left lateral movement have, surprisingly, not hitherto been emphasized in the geologic literature. However, Wiese and Fine (1950, p. 1652) report at least 2 miles of left lateral offset on the Pinon Hill fault (a subparallel branch of the Garlock on the south side of the Tehachapi Mountains), and D. F. Hewett (1950) describes new evidence for recent offset on the Garlock fault zone east of Garlock Station. Thus the sense and recency of movement on this great fault are considered firmly established. The time of initiation and the total amount of left lateral offset on the Garlock fault are not yet known.

Big Pine Fault

Nelson (1925, p. 379–380) first mapped and described the Big Pine fault as it occurs on the south side of Big Pine Mountain. He sketchily mapped the fault from here some 15 miles eastward. Later, Reed and Hollister (1936, p. 94–96) extended the fault to the southeast, as it is likewise shown on the Geologic map of California (Jenkins, 1938). The present mapping shows that this southeastward extension is in error, but that the fault does extend in a clearly revealed trace northeastward to the San Andreas (Pl. 2, fig. 1). Now it is evident that Nelson's east-west portion of the fault and this

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Plate 3. GARLOCK FAULT

Figure 1. Left Lateral Offset of Drainage Lines
Seven miles northeast of Garlock Station.

Figure 2. Left Lateral Offset of Drainage Lines
Ten miles northeast of Garlock Station.

newly mapped and equally long northeast-trending portion gives the Big Pine fault a length of 50 miles and establishes it as a major fault of the region.

The extended portion of the fault trends N. 65°E., with steep dips varying from northwest to southeast and with the throw apparently reversing along its course. At its type locality,[2] 9 miles southwest of Mt. Pinos, the fault separates Miocene continental beds from Eocene marine strata to the south (Pl. 4). Here the fault strikes N. 60°E. and dips 70°S. Striations have been found on slickensided planes of the fault zone which are 70° clockwise from the direction of dip.

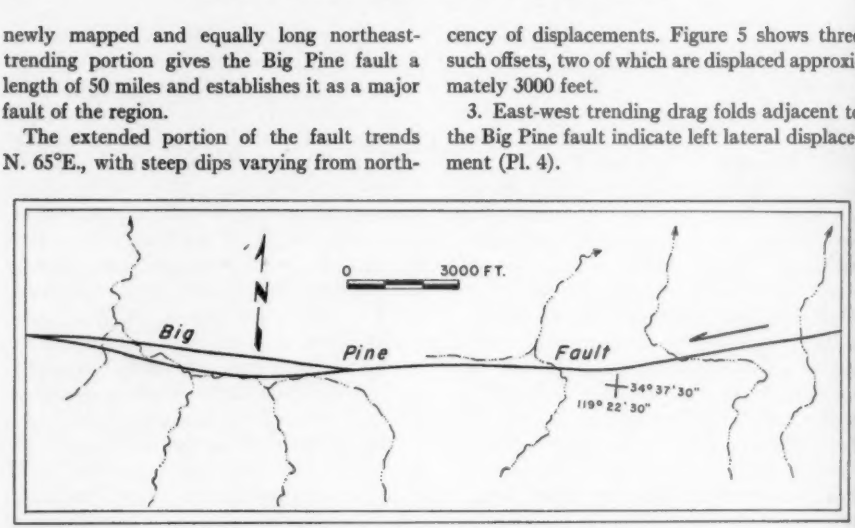

FIGURE 5.—LEFT LATERAL OFFSET OF DRAINAGE LINES BY THE BIG PINE FAULT
Morro Hill and Reyes Peak quadrangles

This left lateral fault has a prominent strike-slip component of displacement as indicated by the following field observations:

1. The oblique-slip striations in the fault zone indicate a predominant strike-slip component of displacement and their orientation combined with the orientation of apparent throw indicates that the southeast side has moved northeastward and up the 70-degree fault plane. On this evidence alone the fault should be classified as a "left lateral reverse," but, since reversals of dip and apparent reversals of throw occur along the fault trace, the appearance of reverse faulting at the type locality is not significant.

2. Left lateral offsets of drainage lines along the Big Pine fault indicate the sense and recency of displacements. Figure 5 shows three such offsets, two of which are displaced approximately 3000 feet.

3. East-west trending drag folds adjacent to the Big Pine fault indicate left lateral displacement (Pl. 4).

4. Left lateral offset of approximately 8 miles since Miocene time is suggested by the possible displacement of a major syncline and a northwest-trending fault contact between Eocene and Miocene strata (Pl. 4).

5. The continental facies of the Miocene, together with its fanglomeratic members extends farther west on the north side of the Big Pine fault than on the south side (Pl. 4).

6. A well 14 miles southwest of Mt. Pinos and just north of the fault (Pl. 4) encountered continental Miocene conglomerate on granitic basement which matches a similar sequence on the south side of the fault 14 miles to the east.

The time of initiation of the Big Pine fault is unknown, but its age must be substantial, considering the probable minimum cumulative displacement of approximately 8 miles.

RELATIONSHIPS OF THE FAULTS

The more important aspects of the San Andreas, Garlock, and Big Pine faults, for the present analysis, are as follows:

1. The faults are essentially contemporaneous, as attested especially by the recency of their displacements.

2. The faults are major structures of the re-

[2] A type locality is picked as recommended by Buddenhagen *et al.* (1930).

gion, as attested by their great lengths and probable great cumulative displacements.

3. Of the three fault zones, the San Andreas is dominant, and its trend predominates in the regional faulting. However, the Big Pine and subparallel Santa Ynez faults separate the northwest-trending Coast Ranges from the east-west trending Transverse Ranges, and the Garlock fault separates the Basin and Range region from the Mojave Desert province.

4. The Big Pine and Garlock faults are on the same trend and are alike in the character of their displacements, as attested by the evidence for left lateral movement on each.

5. The San Andreas right lateral fault zone intersects, at a substantial angle, the Garlock-Big Pine left lateral fault zone.

Based on these data, the more important relationships between the faults appear to be as follows:

1. The Big Pine fault is probably the offset extension of the Garlock fault since its essential characteristics are the same as the Garlock, since it is the only major left lateral fault in the area, and since it abuts the San Andreas at a point northwest of the west end of the Garlock fault, as is consistent with the right lateral movement on the San Andreas. Nolan (1943) expressed the opinion that the Garlock fault should be found about where the Big Pine fault is now mapped but to his knowledge it did not occur. He concluded that, if the Garlock fault did exist west of the San Andreas, it might be offset to the southeast and if so would indicate an anomalous (left lateral) movement on the San Andreas. Finding what was reasonably expected has fortunately then eliminated the possibility of an unlikely reversal of movement on the San Andreas. Furthermore, this shift of 5 miles is in accord with the 5-or 6-mile offset of terrace deposits by the fault 50 miles to the southeast (Wallace, 1949).

2. The big bend in the trace of the San Andreas, that east-west trending portion (Pl. 1), occurs where abutted by the Garlock and Big Pine faults. This anomalous trend is now readily explained by the left lateral movement on the Garlock-Big Pine fault zone. Even the total amount of roughly 25 miles of offset due to the bend is not, in the opinion of the writers, out of line with the possible amount of left lateral offset on the Garlock-Big Pine zone.

3. Another big bend of the San Andreas occurs in the San Bernardino Mountains, approximately 100 miles southeast of the bend described above (Pl. 1). Here possible left lateral movement on the "Warrens Well" fault may be responsible for this other anomalous trend of the San Andreas fault. Critical work on the "Warrens Well" fault would quite likely establish the merit of this inference.

4. Another speculation is that the San Gabriel fault (Pl. 1) is an ancestral portion of the San Andreas. This fault, which occurs south of the Big Pine, extends southeastward on trend with that portion of the San Andreas which lies north of the Big Pine fault. Thus the San Gabriel fault may have been the southeastern continuation of the San Andreas before the big bend was developed. The fact that, according to Crowell (1950), the San Gabriel fault shows evidence of less recent activity than the present San Andreas is a reasonable consequence of such a possibility. Along this line of thinking, the San Jacinto fault (Pl. 1) could be conceived of as also being an earlier representative of the San Andreas in this southern area. Thus the San Jacinto[3] fault would be younger than the San Gabriel fault and perhaps older (although still active) than the present trace of the San Andreas. Likewise the Calaveras and Hayward faults might perhaps be the earlier initiated manifestations of the San Andreas fault zone in the San Francisco Bay region (Pl. 1). It is further suggested that the cumulative right lateral displacement on this greater San Andreas fault system could be of a magnitude of several hundreds of miles. This concept is strengthened by Crowell's report (1952) of 15 to 25 miles of right lateral movement on the San Gabriel fault alone, and during a short interval of geologic time (between late Miocene and late Pliocene). Such a possibility is obviously shocking, particularly when considering its effect on paleogeographic maps which straddle this zone.

Substantiation of the interpretation that the Big Pine and Garlock faults comprise one fault zone is not necessary to the probability that the bend of the San Andreas is related to their left lateral movements or vital to the following

[3] Five miles of right lateral movement is indicated on the San Jacinto fault (Arnett, 1949), and the localization of Jurassic (?) alaskite dikes in the fault zone indicates its early origin (Sprotte, 1949).

analysis of the mechanics of the regional deformation. It is, however, essential that the Big Pine and Garlock once comprised one continuous fault if used as evidence of 5 miles of right lateral offset on the San Andreas. However, since the sense and magnitude of the displacement on the San Andreas from other evidence is compatible with this offset of the Garlock-Big Pine trend, the probability is that a continuous line of faulting actually did at one time exist. Also, the interpretations concerning the other big bend of the San Andreas or the possible historical connections between either the San Gabriel or San Jacinto faults and the San Andreas are not essential to the primary theses of this presentation.

Mechanics of the Faulting

Introduction

The geomorphology of California ranges from high precipitous mountains to low plains (Pl. 1). The larger so-called valleys are usually structural depressions (basins), and many are separated from the mountains by important faults. The geomorphic provinces are also usually geologic provinces, with the main mountain ranges characterized by highly deformed older sediments, metamorphic and plutonic rocks, and the basins characterized by thick, younger and less deformed sedimentary sections (Pl. 1, insert map). Although there are many exceptions, owing in part to the intense and wide-spread late Pleistocene tectonic activity, some of these geomorphic-geologic provinces have persisted through at least most of the Tertiary period, and therefore some of the bounding faults are probably of considerable age.

The principal rock types of California, classified according to characteristic response to deformational stresses, are as follows:

1. Sierran basement complex (pre-Cretaceous): metasedimentary and metavolcanic rocks, intensely deformed and widely invaded by granitic rocks. Because of physical similarity, the Santa Lucia granitics and metamorphics of the southern Coast Ranges and the complexes of the Transverse and Peninsular ranges belong in this group. These are relatively rigid rocks which fail locally by fracturing and, since they or rocks like them are extensively exposed and are presumably of state-wide occurrence at depth, their mechanical behavior is tectonically important.

2. Franciscan basement (pre-Cretaceous): sedimentary and volcanic rocks, regionally unmetamorphosed but highly indurated, commonly intruded by basic igneous rocks which are usually altered to serpentine and have caused local metamorphism. These rocks are exposed in large areas in the Coast Ranges; on the northeast side of the San Andreas fault, and also on the west side of the Nacimiento fault zone. They presumably underlie a much greater area but are probably in turn underlain by granitic rocks. The Franciscan, unlike the granitic basement, is typically incompetent. Although in places intensely fractured, often before being covered by later Jurassic or Cretaceous strata, and usually in fault contact with the other principal rock types, its response to deformational forces has been characterized by folding.

3. Cretaceous and Cenozoic sedimentary and volcanic formations: mainly marine clastic sediments with local volcanics and nonmarine deposits, not strongly lithified and of extremely variable thicknesses and facies. Deposited in large and small basins; locally highly deformed, especially during the late Pliocene-Pleistocene revolution in the Coast and Transverse Ranges, and in uplifts in the Mojave Desert and Salton Sea regions. These rocks form a pliable mantle on the above-described complexes and have therefore responded to tectonic forces primarily by folding, particularly where the sedimentary section is thick or where underlain by Franciscan basement.

The San Andreas, Garlock-Big Pine, and other similar faults are obviously at odds with the geomorphic-geologic provinces and the structural characteristics of the rock types of the region. The unique nature of these faults is shown by their transection of the geomorphic-geologic provinces and by their transection of all the principal rock types, without significant variations.

There are many important northwest-trending steep faults in California which are approximately parallel to the San Andreas, and which are probably also characterized by major right lateral components of displacement (Pl. 1). Several of these are present in the northern

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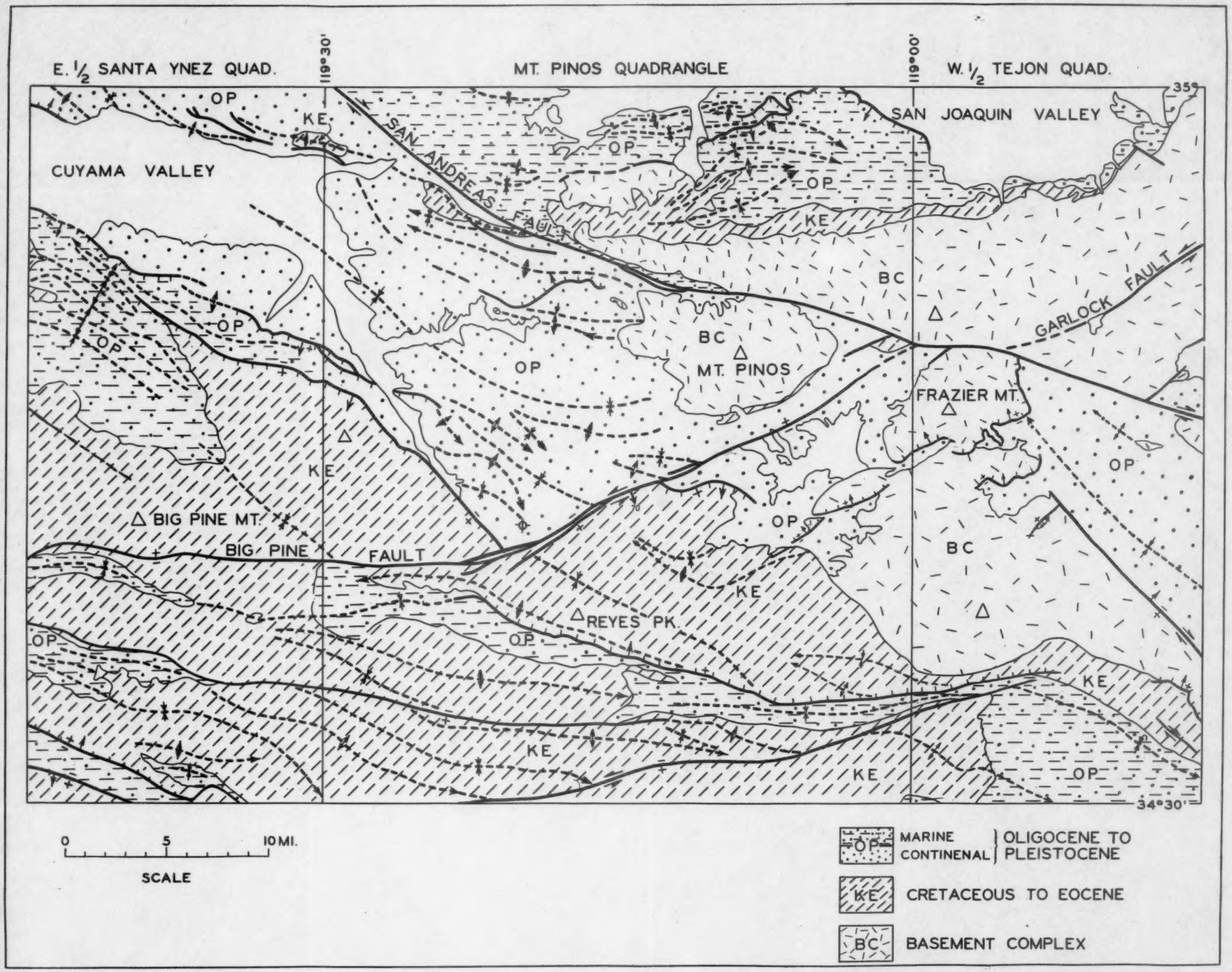

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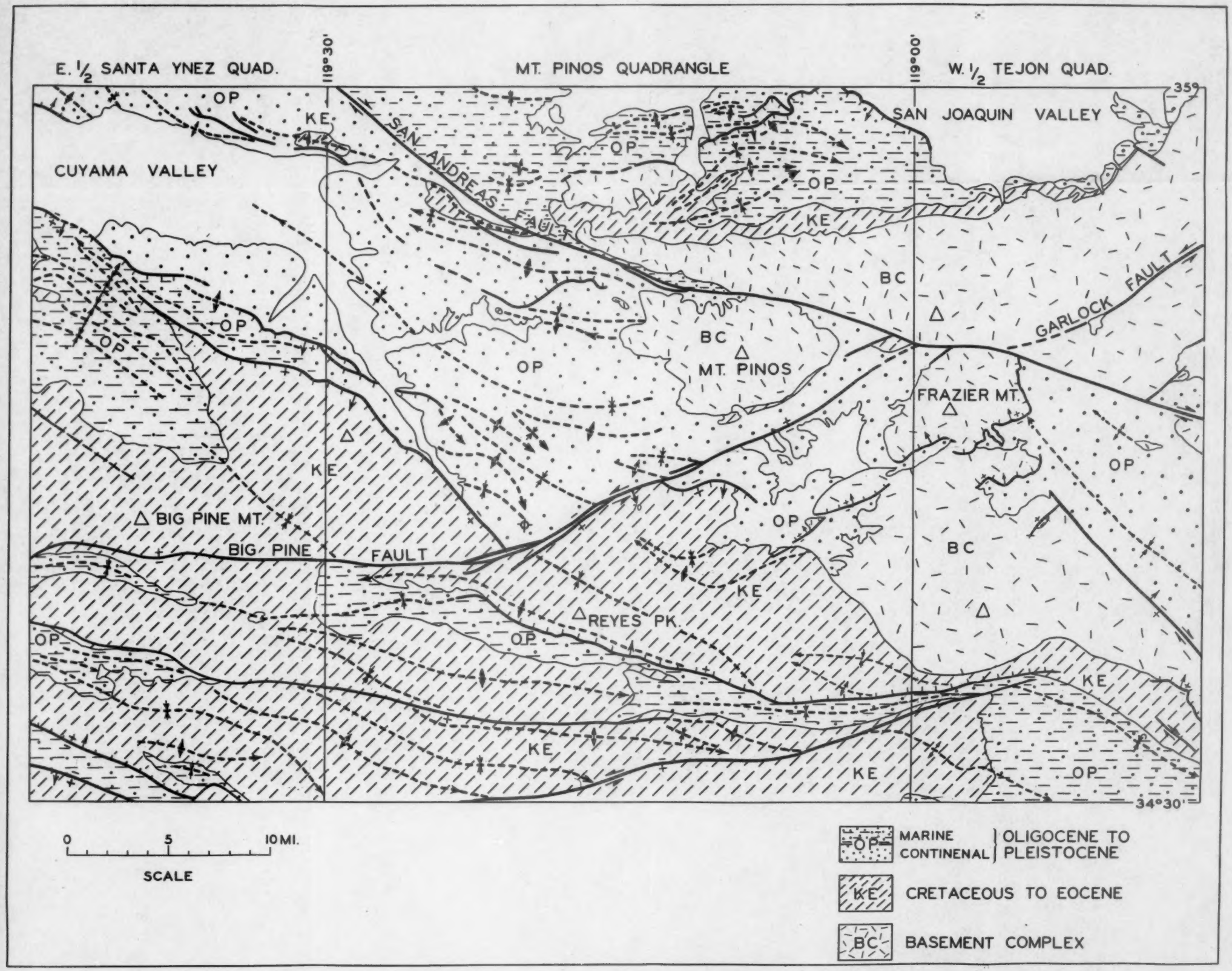

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Coast Ranges, but they have not been described in detail. In the southern Coast Ranges, they include the Calaveras, Hayward, Pilarcitos, San Gregorio, San Marcos, Reliz Canyon, San Juan, and Suey faults, and the long Nacimiento fault zone; in the Transverse Ranges, the San Gabriel and San Jacinto faults; in the Los Angeles Basin, the Whittier and Inglewood faults; and, in the Peninsular Ranges, the San Jacinto and Elsinore faults. The several northwest-trending faults in the Mojave Desert and in the Basin and Range province, some of which are known to affect right lateral offsets, probably also belong in this group. The faults of this set are much more common than the east-northeast set of lateral faults.

The important east-northeast trending lateral faults which, like the Garlock-Big Pine zone, probably have significant left lateral displacements, include the Santa Ynez, White Wolf, possibly the Malibu Beach, faults in the northwestern portion of the San Gabriel Mountains, the "Warrens Well," and faults on Santa Rosa and Santa Cruz Islands. The faults of this set are most common in the Transverse Ranges. Other faults of both sets are quite likely present on the continental shelf.

Confusion should not exist, although it surely does, between these sets of steep lateral faults and those many other faults, both large and small, which indicate vertical relief or local adjustments due to secondary or superficially acting forces.

Large west-northwest trending thrust and reverse faults are especially common in the Transverse Ranges but they can, perhaps usually, be differentiated from the east-northeastward trending Garlock-Big Pine set by degree of dip and orientation of trend. Further confusion will undoubtedly likewise continue to exist between the thrust and reverse faults of the Coast Ranges and the San Andreas type faults. This is particularly true because superficially and locally acting forces are certain to modify the dips and strikes of the steep northwest-trending right lateral faults.

Strong folding of Cretaceous and Cenozoic strata is very common in both the Coast and Transverse ranges. In the Coast Ranges, the main folds and associated faults typically trend a little westward from the northwest San Andreas trend. In the Transverse Ranges, the main fold axes typically trend a little north of west in contrast to the east-northeast trend of the Garlock-Big Pine set and again the major thrust and reverse faults are approximately parallel to the folds.

It is here hypothesized that the folds and accompanying thrust and reverse faults of the Coast and Transverse ranges are shallow structures compared to the steep, and surely deep, lateral fault system. This concept furthermore suggests that these folds and faults which affect upward relief are secondarily related to the deep-seated forces which presumably developed the lateral fault system of nearly horizontal relief. Other minor faults and folds of diverse trends are even less directly related to the primary regional forces. As an example, the Kettleman Hills anticline of the southern Coast Ranges is a minor structure compared to the San Andreas fault zone, but the faults on it are still more local and superficial and thus more remotely related to the regional tectonic forces.

To ascertain the orientation of deformational forces which develop even minor structures is fraught with difficulties, but in such analyses, based on good mapping and sound principles of mechanics, lies the possibility of taking the Coast Ranges of California out of the defeatist category of a "heterogeneous mobile belt".

Strain Pattern

Since the Garlock-Big Pine and San Andreas fault zones are the major fracture features of the region, and since they are concurrently active, they are probably genetically related. Then since the displacements of these intersecting fault zones show a systematic relative east-west extension and north-south shortening, their common origin seems to be an inevitable conclusion. Therefore, these fault zones are considered to comprise a "strain pattern" or system which is the result of a particular set of deforming forces.

This strain pattern is considered to be the primary fracture system of the entire region because these are the long faults (measured in hundreds of miles); they are deep faults (at least 10 miles as indicated by the usual calculated depth of foci of earthquakes in this region); and they are probably the faults of

maximum cumulative displacement. Other large faults such as the San Gabriel, Nacimiento, and Santa Ynez (Pl. 1) may be nearly as directly related to the primary regional stresses, but the multitude of smaller faults and folds are undoubtedly in secondary or

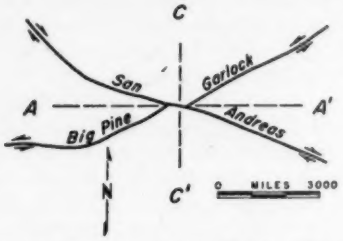

FIGURE 6.—SAN ANDREAS AND BIG PINE-GARLOCK STRAIN SYSTEM
AA′ axis of relative extension; CC′ axis of maximum shortening

tertiary order strain systems which are only indirectly related to the primary stress pattern.

The diagram (Fig. 6) of the northwest-trending San Andreas right lateral shear zone and the northeast-trending Garlock-Big Pine left lateral shear zone shows the CC_1 axis of shortening and AA_1 axis of relative extension. This strain pattern is geotectonically important because its size of approximately 600 x 200 miles makes it a respectable sample of the earth's crust. Even this area of systematic deformation may be, however, only a portion of a much larger unit or even of secondary or tertiary order in relation to the primary strain systems of the crust. The amount of deformation since the inception of this shear system is probably also geotectonically important, but no conclusive evidence is now known by which to date the first displacements or determine the total cumulative movement on either of the fault zones.

Supposing, however, as now seems tenable, that the right lateral displacement on the San Andreas amounts to several hundred miles and the left lateral displacement on the Garlock-Big Pine zone totals several tens of miles, some of the consequences should be as follows: (1) The San Andreas was initiated first and much movement occurred on it before the inception of the Garlock-Big Pine fault; (2) After the Garlock-Big Pine was established, it caused the big bend in the San Andreas and was offset by the San Andreas; and (3) The Garlock-Big Pine zone interrupted the continuity of displacement on the San Andreas and thereafter the north-south wedges actually moved toward each other.

If one side of the San Andreas has actually done most of the moving, indicated by the progressive offsets of Figures 2 and 3, there now appears to be a basis for determining which was the mobile block. Thus after the inception of the Garlock-Big Pine zone and assuming crustal shortening, the north-south wedges have been moving toward each other. Therefore it seems probable that the east side of the San Andreas north of the Garlock fault has actually been moving, as has the west side south of the Big Pine fault (Fig. 6).

Stress Pattern

The character and orientation of the forces responsible for the primary strain pattern of this large region are not readily definable. Even in this fortunate case where the fault characteristics and relationships clearly show a pattern of deformation, there are many orientations of deforming forces which could cause the observed deformation. An example in which the character and orientation of deforming forces are more obvious is in the case of "feather joints" (Cloos, 1932) where the fractures can be related to secondary forces which arise from known movements. In the case of the subject region, however, we are dealing with the major known deformation and thus cannot know the character and orientation of the causal forces. On the other hand, by knowing the deformational pattern of this region, it may be possible to resolve its relationship to a larger strain pattern which may in turn indicate the stress pattern in this region.

At present, however, only general deductions may be made. The first of these is that the deformation is probably the result of compressive stresses because: (1) experimental work shows that rocks are brittle enough so that tensional stresses would be apt to cause tensional fractures instead of this shear pattern, and (2) regionally important thrust and reverse faults and folds show prominent crustal shortening (and prominent upward relief with the AA_1 axis of extension locally oriented vertically). Thus, assuming compressive stresses,

the possible effective orientations (Fig. 7) are north-south oppositional forces, or a northeast-southwest counterclockwise couple, or a northwest-southeast clockwise couple. The northeast-southwest counterclockwise couple is favored

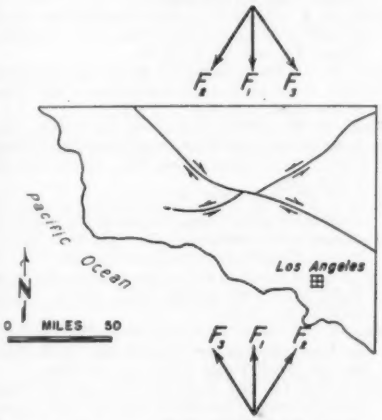

FIGURE 7.—STRESS SYSTEMS FOR SAN ANDREAS AND BIG PINE-GARLOCK FAULTS

F_1 direct compression; F_2 counterclockwise rotational compression; F_3 clockwise rotational compression.

because this orientation could perhaps more logically cause both the prominent northwest-trending folds of the Coast Ranges and the east-west folds and reverse faults of the Transverse Ranges (by upward relief in combination with the north-south shortening of the major shear pattern). With directly acting north-south compression, a more nearly equal manifestation of the shear sets would be expected, whereas the northwest right lateral set greatly predominates in this region. Likewise a northwest-southeast clockwise couple is unlikely because, where local upward relief occurred, the folds would be likely to trend northeast-southwest, whereas the general trend of folds in the Coast Ranges is northwestward.

GENETIC CONCEPTS AND CONCLUSIONS

Because the San Andreas and Garlock-Big Pine fault zones are major structures of great length, depth, displacement, and age; because they constitute a strain pattern of north-south shortening; because the probable orientation of causal forces is northeast-southwest; and because these forces are operative at great depth, a drag mechanism due to flowage of rock material from the Pacific basin to the continent of North America may be the primary reason for deformation in this region.

The mechanical effect of subcrustal convection currents, a presently favored concept (Gilluly *et al.*, 1951), may have caused this deformation. Such a drag effect near oceanic margins could, then, be responsible for the permanence and growth of continents, development of geosynclines, deformation of geosynclinal deposits, association of metamorphic and igneous rock facies with orogeny, development of mountain roots, and possibly, the development of major lateral fault systems.

If the major lateral fault system described here is really so directly related to primary forces of deformation, much advancement in the understanding of geotectonics may be expected by the world-wide study of such faults. Perhaps faults with lateral displacements of hundreds of miles could even be responsible for some geologic relationships which seem to require land bridges.

Entirely apart from these speculations, the following conclusions are attained. First, the mapping of the faults and the evidence for the sense of the displacements are considered reliable. Second, the essential contemporaneity of the faults is considered to be proved. Third, this pattern of conjugate shears with opposing centripetal and centrifugal moving wedges is believed to be mechanically sound. Fourth, this strain system shows definite north-south shortening and relative east-west elongation of, during its activity, at least tens of miles and affecting a crustal area of at least 120,000 square miles; as such it is of significant proportions with respect to earth deformation. Fifth, this strain system is, as far as known, compatible with lesser structural features in the region and has in itself set up stress conditions which have developed lower-order strain systems. This major tectonic pattern in California may thus provide a satisfactory framework for the understanding of many elements of deformation within this region. Sixth, if deformation in other large areas proves to be compatible, it may ultimately be possible to develop a system of primary geotectonic forces. Hopefully this approach could, in conjunction with

other lines of research, culminate in a usable theory of the causes of earth deformation.

Before the geologic history of California is satisfactorily established, much more critical mapping is needed, especially to determine the sense, duration, and amount of lateral movements on faults. Displacements which are measured in miles will have significant effects on paleogeologic and paleogeographic maps, and even those which are measured only in thousands of feet may, for example, be very critical in studies of the distribution of sedimentary facies. As these data accumulate, we can expect to attain a more nearly complete story of the dynamic geologic history of the region.

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TERTIARY GEOLOGY OF THE GALISTEO-TONQUE AREA, NEW MEXICO

BY CHARLES E. STEARNS

ABSTRACT

The Galisteo-Tonque area lies on the east margin of the Rio Grande Depression, in north-central New Mexico. It extends from the north end of the Sandia Mountains to the south end of the Sangre de Cristo Mountains.

The Eocene Galisteo formation, deposited by streams flowing southwest from the Sangre de Cristo Mountains, was buried by volcanic detritus deposited in great alluvial fans peripheral to contemporary eruptive centers in the Ortiz Mountains and Cerrillos Hills. Surface volcanism was accompanied and followed by emplacement of stocks, laccoliths, sills, and dikes, the larger bodies localized in a north-south line, marked by the present Cerrillos Hills, Ortiz, San Pedro, and South Mountains. Igneous activity heralded and accompanied regional deformation, during which broad, northward-plunging, faulted folds were formed in a block north of a northeast-trending fault system.

Oligocene (?) deformation and igneous activity were followed by a period of erosion, interrupted locally by extrusion of limburgite and basalt flows (Cieneguilla limburgite). Subsequent broad warping in the early (?) Miocene led to deposition of alluvial and lacustrine beds in a sedimentary basin centering northwest of the Galisteo-Tonque area, in the Rio Grande Depression. Detritus was contributed locally to the basin from highlands eroded in the areas of Oligocene uplift, and tuffaceous materials in a part of the basin probably represent the distal end of the Abiquiu tuff fan, spread from contemporary centers of eruption in northernmost New Mexico.

Deposition of the Santa Fe formation in the same basin continued during the Miocene and Pliocene. The Abiquiu tuff fan was buried by arkosic fans built southwestward from the Sangre de Cristo Mountains. During the late Pliocene, movement began on some of the normal faults bounding the present Rio Grande Depression. The boundary faults outline a structural depression coincident with the earlier basin of deposition in a general way, but differing considerably in detail. The diastrophic movements were reflected first by lithologic changes in the Santa Fe formation, and eventually by the change from deposition to erosion in the Rio Grande Depression.

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Introduction

The present study began as an investigation of the late Tertiary history of the Rio Grande Depression and its contained deposits, extending the general survey undertaken by the late Kirk Bryan and his students. The Galisteo-Tonque area lies on the margins of the Santo Domingo Valley (Fig. 1), studied in detail by Bryan and Upson (unpublished manuscript), and in the area of offset between the Albuquerque-Belen basin of central New Mexico and the related basin system trending northward from Santa Fe into southern Colorado.

Extensive exposures of the early Tertiary Galisteo formation, enhanced by the discovery of vertebrate fossils in it, offered the promise of elucidating the earlier Tertiary history of an area adjacent to the Rio Grande Depression, thereby implementing comparison of early and late Tertiary stratigraphic and structural histories. Accordingly, our attention was drawn from the Rio Grande Depression proper into the area marginal to it. Descriptions of the early Tertiary sedimentary and igneous rocks (Stearns, 1943; 1952b) are summarized here. Subsequent deformation is described in some detail as are the late Tertiary rocks and their structure.

The readily accessible Galisteo-Tonque area has been the object of considerable reconnaissance and some detailed study. References to early exploration are included in detailed reports on the Cerrillos Hills (Johnson, 1903), on the mining districts of the Cerrillos Hills, Ortiz and San Pedro Mountains (Lindgren, Graton, and Gordon, 1910, p. 163–175), and on the Cretaceous rocks (Lee and Knowlton, 1917, p. 201–217). Excellent summaries and a geologic map were published by Darton (1928a, p. 94–103, 255–266; Pl. 24). Recent studies of the Paleozoic stratigraphy in much of New Mexico (Read and Wood, 1947) have included the Galisteo-Tonque area. Bryan and Upson (unpublished manuscript) included portions of the area in their study of the adjacent Santo Domingo Valley; their unpublished material has been placed at the writer's disposal both in manuscript and in conference with the senior author.

The area was re-mapped on planimetric drainage maps (scale 1 inch to the mile), compiled from aerial photographs by the Soil Conservation Service. Field locations were made on contact prints of the aerial survey, at twice the scale, and transferred to the drainage maps. Stratigraphic sections were measured by pace and compass traverse.

ACKNOWLEDGMENTS

The present report stems from the patient and inspiring guidance of the late Professor Kirk Bryan, of Harvard University. It is not feasible to separate the many ideas herein which are properly his own; his generous discussions in field and office are gratefully remembered. Professor Esper S. Larsen, Jr., of

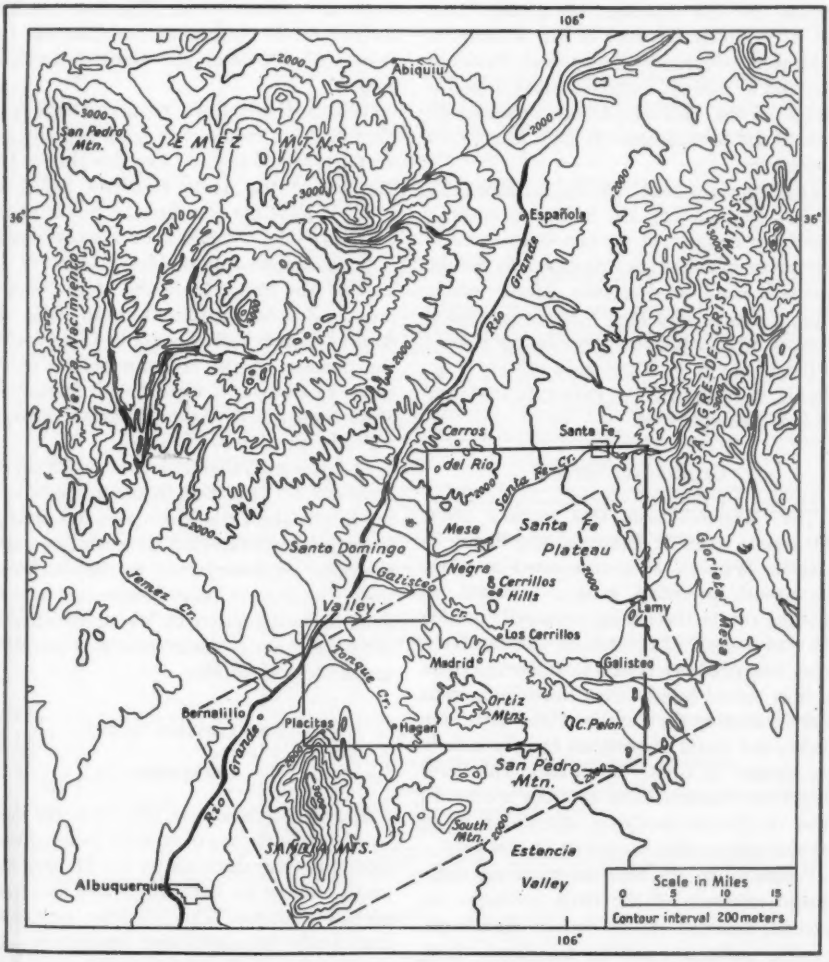

FIGURE 1. TOPOGRAPHIC MAP OF A PART OF NORTH-CENTRAL NEW MEXICO

Redrawn from Darton (1928b). Galisteo-Tonque area (Pl. 1) is outlined by solid line; dashed lines enclose area shown in Figure 7 (Structures of the Pre-Abiquiu (?) rocks).

Harvard University, visited the writer in the field and supervised brief petrographic studies. Dr. John B. Reeside, of the U. S. Geological Survey, kindly surveyed fossils collected from Cretaceous beds. Mr. Edward A. Schmitz drafted the line drawings; the geologic map (Pl. 1) is largely the work of Mr. William N. Palmstrom. Professor Bryan and Professor Marland P. Billings suggested many improvements in the manuscript; Mrs. Charles E. Stearns and Miss Beatrice L. Dole aided in its preparation.

Field expenses were met by the generosity of the writer's father, Dr. A. Warren Stearns. Numerous residents of the area contributed to the success of field work, but especially did the Seligman family of Bernalillo, Mr. Benjamin Cano of Hagan, and Mrs. J. C. Underwood of Underwood Ranch. The writer was ably assisted in the field by A. F. Marston, in 1939; by J. K. Grunig, in 1940; by N. O. Parker, in 1941; and by I. S. Fisher, in 1946.

Geographic Setting

The Galisteo-Tonque area includes about 900 square miles in Sandoval and Santa Fe counties (Fig. 1). It is dominated by three north-south mountain ranges, arranged en échelon, two of them lying principally outside the area mapped. These are the Sandia Mountains, long cited as a typical fault block; a chain of hills carved largely from intrusive igneous rocks, including the Cerrillos Hills, Ortiz, San Pedro, and South Mountains; and the sprawling Sangre de Cristo Mountains. The lesser heights of Glorieta Mesa continue southward from the Sangre de Cristo Mountains, as an eastern topographic boundary to the area.

Peripheral to the highland areas are widespread remnants of the Ortiz pediment, an ancient, dissected erosion surface. Both topographic uniformity and an associated, thick, gravel deposit emphasize the physical discontinuity provided by these remnants between the highland areas and the lowland areas, carved principally by tributaries to Galisteo and Tonque Creeks.

Galisteo Creek enters the area at Lamy, on the boundary between the Sangre de Cristo Mountains and Glorieta Mesa, and flows southwest to the town of Galisteo. The town occupies the center of the broad Galisteo Lowland, excavated in Cretaceous and early Tertiary rocks below the high-level Santa Fe Plateau to the north and Estancia Valley to the south. Cerro Pelon, carved from a thick sill, isolates a smaller lowland to the southwest, the Canyada Las Jollas (Pl. 1), from the Galisteo Lowland proper.

Galisteo Creek flows northwestward from the Galisteo Lowland, through a gap between the Ortiz Mountains and Cerrillos Hills at Los Cerrillos to the Santo Domingo Valley, in which it joins the Rio Grande. North of the Cerrillos Hills, the Santa Fe Plateau merges with the basalt-capped Mesa Negra de La Bajada. The west edge of the Mesa Negra is the La Bajada escarpment, overlooking the Santo Domingo Valley. Santa Fe Creek funnels the drainage of the northern portion of the Santa Fe Plateau, in a canyon cut through the Mesa Negra, to the Santo Domingo Valley, in which it joins the Rio Grande.

The Tonque Valley is eroded in Cretaceous and early Tertiary rocks in the saddle between the Sandia and Ortiz Mountains. Its drainage is carried northwestward around a low prong extending northward from the Sandia Mountains. The Tonque Valley is separated from the Santo Domingo Valley by low spurs extending northwest from pediment remnants peripheral to the Ortiz Mountains.

Pre-Tertiary Rocks

Introduction

The pre-Tertiary rocks (Fig. 2) of the Galisteo-Tonque area are of interest to the present study as source materials for the Tertiary sediments and for the light which they shed upon general structure. The following paragraphs draw freely upon published descriptions, supplemented in a few cases by observations by the writer.

Precambrian

The basement complex of the Galisteo-Tonque area is a variety of igneous and metamorphic rocks, generally regarded as Precambrian. These rocks form the rugged west scarp

AGE	MAP SYMBOL	FORMATION NAME	THICKNESS	LITHOLOGIC CHARACTER
UPPER CRETACEOUS	Kmv	MESAVERDE FORMATION	0 to 1000'	Sandstone, shale, and coal
	Kmvc	MANCOS SHALE	800' to 1000'	Sandy shale, abundant large limestone concretions, thin lenses of sandstone.
				CANO MEMBER of Mesaverde formation (sandstone) present only in Tonque Valley.
	Km_3			Sandy shale, abundant limestone concretions. Thin lenses of sandstone near top
				Paper-thin sandstone and sandy shale.
	Km_2		500'±	Dark shale
	Km_1		600' to 800'	JUANO LOPEZ MEMBER: Dark, thin-bedded limestone.
				CARLILE SH.: Dark shale, locally prominent sandstone beds.
				GREENHORN L.S.: Platey dark limestone, weathers gray.
				GRANEROS SH.: Dark shale, locally prominent sandstone beds.
				DAKOTA (?) FM.: Discontinuous sandstone, carbonaceous shale.
JURASSIC	J	MORRISON FORMATION	500'±	Buff to white sandstone; intercalated variegated shale
		WANAKAH FM.	125'-250'	Basal limestone; variegated shale
		ENTRADA S.S.	30'-120'	Buff, pink, and white sandstone
TRIASSIC	Tr	DOCKUM GROUP	800'±	Red and variegated sandstone and shale
PERMIAN	Cy	GLORIETA S.S.	150'±	Yellow calcareous sandstone, local conglomerate and limestone
	Cay	YESO FORMATION	400'±	Light red sandstone, siltstone, and shale
		ABO FORMATION	1100'±	Sandstone, arkose, and shale, Predominantly red in color
PENNSYLVANIAN	Cm	MAGDALENA GROUP	550'±	MADERA LIMESTONE: "Arkosic limestone" member Limestone and yellow or green sandstone
			450'±	MADERA LIMESTONE: "Gray limestone" member Massive limestone, abundant chert
			125'	Sandia formation: sandstone, shale, subordinate limestone
PRE-CAMBRIAN?			0-125'	Unnamed unit: massive limestone with clastic beds at base
PRE-CAMBRIAN	p-€			Quartzite, gneiss, schist, granite

FIGURE 2. STRATIGRAPHIC COLUMN OF PRE-TERTIARY ROCKS

of the Sandia Mountains (Ellis, 1922, p. 21–29). Distinctive among them are massive, bluish-white quartzite and coarse-grained, pink or gray granite. The latter is probably the youngest member of the complex. Similar rocks crop out south of the Galisteo-Tonque area, on the gentle east slope of the Sandia Mountains, and in a broad area west of South Mountain (Fig. 1).

Precambrian rocks form most of the foothills of the Sangre de Cristo Mountains in the northeast portion of the area. Here, pink or red

granite-gneiss is abundant, but schists of various types occur.

Pennsylvanian Series

Magdalena group.—The Precambrian crystalline rocks of the Sandia Mountains are overlain with strong angular unconformity by Pennsylvanian marine strata assigned to the Magdalena group. The following summary draws freely from a recent report by Read and Wood (1947).

An unnamed unit, comprising 16 feet of green quartz conglomerate and 90 feet of massive limestone, occurs locally at the base of the Magdalena group in the northern part of the Sandia Mountains. Fossils have not been found in it, and it may be of pre-Pennsylvanian age.

The Magdalena group proper includes the Sandia formation and the Madera limestone. They were not mapped separately by the writer, but most of the area shown as the Magdalena group in Plate 1 is underlain by the Madera limestone. The older Sandia formation is principally sandstone and shale, with subordinate limestone and conglomerate. Thickness is variable; east of Placitas it is about 125 feet. The younger Madera limestone, chiefly massive limestone, is about 1000 feet thick. Two members are recognized. The lower ("gray limestone") member is dominantly limestone, with abundant chert both as nodules and as thin beds. Beds of yellow or green sandstone are interbedded with limestones in the upper ("arkosic limestone") member. On the east slope of the Sandia Mountains, red shale is also common in the upper member, but it was observed at only a few localities in the Galisteo-Tonque area. The Madera limestone contains Pennsylvanian fossils (Needham, 1937, p. 12; Thompson, 1942, Plate 1).

Scattered outcrops of limestone along the base of the Sangre de Cristo Mountains between Lamy and Santa Fe are presumably equivalent to part of the Magdalena group.

Permian Series

Permian rocks crop out near Placitas in a belt stretching south from Alamos Altos ranch along the east base of the Sandia Mountains, turning eastward at Canyon Benado. On the geologic map (Pl. 1) rocks assigned to the Abo and Yeso formations are not separated. Equivalent beds northeast of Lamy, now assigned to the Sangre de Cristo and Yeso formations (Read *et al.*, 1944), are also mapped as combined Abo and Yeso. The overlying Glorieta sandstone, a fine marker, is shown for the purpose of bringing out structural relations.

ABO FORMATION: The Madera limestone is overlain conformably by the Abo formation, comprising chiefly sandstone, arkose, and shale, predominantly red. The contact between the two formations is transitional, and thin beds of limestone occur in the base of the Abo formation. A few beds of limestone conglomerate were observed near the base of the formation. The limestone conglomerate and arkose imply that both Precambrian rocks and the Magdalena group were exposed to erosion during Abo time (Lee, 1909, p. 13). The limestone conglomerate contains pebbles as much as 2 inches in diameter, and at one locality near Canyon Benado, an arkose consists of angular, granitic fragments half an inch across. It is believed, however, that these beds record erosion outside the areas mapped.

The Abo formation is approximately 1000 feet thick at the north end of the Sandia Mountains (Read *et al.*, 1944, sec. 16); it is probably of Permian age, but the Pennsylvanian-Permian boundary has not been precisely established in this area.

YESO FORMATION: About 400 feet of light-red sandstone and siltstone, with a few thin beds of limestone, represent the Yeso formation (Read *et al.*, 1944). These beds rest with apparent conformity upon the Abo formation.

GLORIETA SANDSTONE: The Yeso formation is overlain with apparent conformity by about 150 feet of yellow sandstone. Lenses of conglomerate, containing quartzite pebbles 1–3 inches in diameter, are common. The beds are firmly cemented with calcium carbonate, and the sandstone is generally a prominent ridge-former. Thin beds of sandy limestone occur locally. The Glorieta sandstone is generally correlated with sandstones in the base of the San Andres limestone (King, 1942, p. 688).

Triassic System

Dockum group. The Glorieta sandstone is overlain by several hundred feet of sandstone and shale, generally poorly exposed. The beds are principally brick-red, but dark-blue, purple, brown, green, yellow, and white beds also occur. These rocks underlie part of the Placitas area and a wide belt on the west and south sides of the Tonque Valley. They are further exposed in tributaries of Galisteo Creek north and south of Rosario Siding, and extensively in Glorieta Mesa. Upper Triassic fossils are reported from the latter locality (Romer, 1939). These beds are generally correlated with the Dockum group of eastern New Mexico.

Jurassic System

Jurassic rocks, mapped as a single unit (Pl. 1), crop out near Placitas, along the east and south sides of the Tonque Valley, in the Galisteo Monocline north and south of Rosario Siding, at scattered localities on the pediment west of the Ortiz Mountains, and on the west flank of Glorieta Mesa. The Jurassic rocks of central New Mexico, including those of the Galisteo-Tonque area, have been variously correlated with those of other areas (Darton, 1928a; Baker, Dane, and Reeside, 1936; 1947; Heaton, 1939; Goldman and Spencer, 1941). The present paper follows the latest correlation (Baker, Dane, and Reeside, 1947).

ENTRADA SANDSTONE: The red beds of the Dockum group are overlain by 30–120 feet of massive sandstone, generally bipartite. The lower part is fine-grained and red, pink, or pink and white. Its thickness varies, and in the southern Tonque Valley it is apparently absent. The upper part is medium-grained and more uniformly buff in color. It contains numerous rounded and frosted grains of quartz, at least locally. It is universally present, and of rather uniform thickness (30–50 feet). The uppermost 2 or 3 feet of the buff sandstone are thin-bedded and grade into the overlying Todilto limestone. A similar, bipartite division of the Entrada sandstone is reported in adjacent areas (Renick, 1931, p. 30; Heaton, 1939, Fig. 13; Harley, 1940, p. 22). The Dockum group and Entrada sandstone appear to be conformable.

WANAKAH FORMATION: The Entrada sandstone is overlain by the distinctive dark-gray, very thin-bedded, fetid, Todilto limestone, a member of the Wanakah formation. It is 5–15 feet thick in the Galisteo-Tonque area. In many localities, bedding planes are intricately contorted. Near Tonque and in the Galisteo Monocline, the limestone is overlain by 40–100 feet of massive gypsum. In the southern part of the Tonque Valley, however, and on the east side of the Galisteo Lowland, massive gypsum is absent. In these areas, the upper part of the limestone is massive rather than thin-bedded. The Todilto limestone and overlying gypsum are probably equivalent to the Curtis formation of Utah.

Immediately overlying the gypsum are 100–150 feet of maroon shale, containing a few lenses of buff or white sandstone. At least in part, these beds are presumably also equivalent to the Wanakah formation of southwestern Colorado.

MORRISON (?) FORMATION: The Wanakah formation is succeeded by 400–600 feet of thick sandstones—gray, white, or buff—with interbedded shales of various hue. Locally, the sandstones are pebbly or conglomeratic. In the Tonque Valley, numerous pebble lenses contain highly polished, varicolored, siliceous pebbles. The pebbles apparently correspond to the "gastroliths" of the Morrison formation in other areas, although an origin independent of dinosaurian digestion seems probable (Stokes, 1942). Entirely similar pebbles occur in the early Tertiary Galisteo formation of the Galisteo-Tonque area (Stearns, 1943, p. 305). On the east side of the Galisteo Lowland, conglomerate lenses are numerous and coarse. The contained pebbles are of siliceous types, and in many cases subrounded to subangular.

The dominantly sandstone beds overlying the Wanakah are provisionally assigned to the Morrison (?) formation, but their equivalence to the Morrison formation of other areas is by no means established. It is entirely possible that they are equivalent in part to the Purgatoire formation (Lower Cretaceous) of eastern New Mexico.

Upper Cretaceous Series

Upper Cretaceous rocks underlie most of the Galisteo Lowland, the Galisteo Monocline west of Los Cerrillos, the Tonque Valley, and the Placitas area. They comprise two broad, generally recognizable, units, a lower section of shale, with subordinate sandstone and limestone, commonly called the Mancos shale, and an upper section of sandstone, shale, and coal commonly called the Mesaverde formation. In addition, sandstones at the base of the Mancos shale are commonly separated as the Dakota (?) formation, although precise equivalence of the beds so named has not been established. These rocks have been described in some detail (Lee and Knowlton, 1917, p. 201–217; Rankin, 1944; Stearns, 1952a).

In the present study, three subdivisions of the Mancos shale were mapped to illustrate the general structure. The first subdivision includes the several formations recognized by Rankin below the Juana Lopez member of the Carlile shale, a fine horizon-marker (Rankin, 1944, p. 26). The Dakota (?) formation and Greenhorn limestone are also mapped separately in the Galisteo Monocline and the Galisteo Lowland. The boundary between the second and third subdivisions is a striking lithologic change within beds of Niobrara age from dominantly argillaceous shale to dominantly arenaceous shale and thin sandstone.

DAKOTA (?) FORMATION: The Dakota (?) formation in the southern part of the Tonque Valley comprises 6–25 feet of well-bedded, brown, siliceous sandstone, forming a prominent hog-back. In the Galisteo Monocline and on the east side of the Galisteo Lowland, carbonaceous shales interbedded with gray-white sandstone at the base of the Mancos shale were mapped as Dakota (?). However, comparable beds are not present in many localities, and mapping the formation as a continuous unit in the Galisteo-Tonque area is a generalization not consistent with field data.

GRANEROS SHALE: The Graneros shale comprises 200–250 feet of dark shale and subordinate sandstone. In most localities, a brown, thin-bedded, siliceous sandstone 5–30 feet thick (Tres Hermanos sandstone) forms a prominent hog-back at about the middle of the section.

GREENHORN LIMESTONE: The Greenhorn limestone is represented by 30–40 feet of interbedded limestone and calcareous shale. As many as twelve thin limestone beds occur in the Galisteo Lowland; no more than one bed is persistent in the Tonque Valley.

CARLILE AND NIOBRARA SHALES: The great thickness of shale lying between the Greenhorn limestone and the Mesaverde formation is principally equivalent to the Carlile and Niobrara shales of standard nomenclature. The shales are poorly exposed; a few thin but persistent sandstone beds serve as local horizon markers. The Carlile-Niobrara contact is not marked by a mappable lithologic horizon, but the Juana Lopez member of the Carlile shale is a close approximation to it (Rankin, 1944, p. 26). This member, about 6 feet of interbedded arenaceous shale and very thin-bedded, arenaceous, fetid limestone, forms a low but very persistent hog-back 300–475 feet above the Greenhorn limestone and marks the break between "lower Mancos shale" and "middle Mancos shale" in Plate 1.

The Juana Lopez member is overlain by about 500 feet of argillaceous shale ("middle Mancos shale"). A succeeding zone of sandy shale and paper-thin sandstone, commonly exposed in a rounded hog-back or series of hog-backs, was chosen as the base of the "upper Mancos shale" (Pl. 1). This zone is 250–300 feet thick, and is overlain by 800–1000 feet of more or less sandy shale. Thin lenses of sandstone are common in the upper part of the shales in the Tonque Valley, and the Cano sandstone member of the Mesaverde formation is contained within them.

MESAVERDE FORMATION: The Mesaverde formation comprises massive buff-white to brown sandstone, brown or black carbonaceous shale, and coal. In the Tonque Valley, the first massive sandstone characteristic of the formation is the Cano sandstone member (Stearns, 1952a), a tongue interbedded with the upper part of the Mancos shale. Its thickness increases southward from 25 feet in the Arroyo Pinovetito to at least 220 feet southeast of Hagan. It was not recognized in the Galisteo Monocline and the Galisteo Lowland.

The thickness of the upper member of the Mesaverde formation in the Tonque Valley,

and of the entire Mesaverde formation in other parts of the area, varies considerably. Maximum thicknesses of about 1000 feet are preserved near Hagan and near Madrid; in the north half of the Galisteo Lowland, the formation was entirely removed by pre-Galisteo erosion.

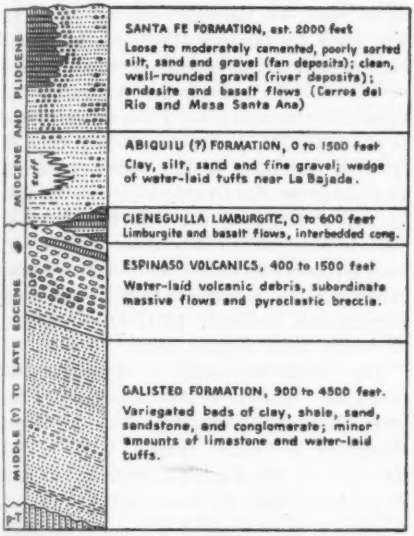
MIOCENE AND PLIOCENE
MIDDLE (?) TO LATE EOCENE
P-T
SANTA FE FORMATION, est. 2000 feet
Loose to moderately cemented, poorly sorted silt, sand and gravel (fan deposits); clean, well-rounded gravel (river deposits); andesite and basalt flows (Cerros del Rio and Mesa Santa Ana)
ABIQUIU (?) FORMATION, 0 to 1500 feet
Clay, silt, sand and fine gravel; wedge of water-laid tuffs near La Bajada.
tuff
CIENEGUILLA LIMBURGITE, 0 to 600 feet
Limburgite and basalt flows, interbedded cong.
ESPINASO VOLCANICS, 400 to 1500 feet
Water-laid volcanic debris, subordinate massive flows and pyroclastic breccia.
GALISTEO FORMATION, 900 to 4500 feet.
Variegated beds of clay, shale, sand, sandstone, and conglomerate; minor amounts of limestone and water-laid tuffs.

FIGURE 3. STRATIGRAPHIC COLUMN OF TERTIARY ROCKS

EARLY TERTIARY ROCKS

Introduction

Early Tertiary sedimentary rocks (Fig. 3; Pl. 1) crop out near Placitas, in the Tonque Valley, 3 miles south of Rosario Siding, in the La Bajada escarpment, near La Cienega, and in a broad area between the Cerrillos Hills, Ortiz Mountains, and Lamy. The following brief summaries are taken from earlier descriptions (Stearns, 1943; 1952b).

Galisteo Formation

The Upper Cretaceous rocks of the Galisteo-Tonque area are overlain unconformably by the Galisteo formation of late Eocene age. The Galisteo formation comprises 900 to at least 4000 feet of sandstone, sand, clay, and shale, generally variegated in color, with minor amounts of conglomerate, fresh-water limestone, and water-laid tuff. The beds were deposited by rivers flowing into and through a broad basin, which was deepened and enlarged by warping contemporary to deposition. A persistent horizon of massive conglomerate, containing well-rounded pebbles of limestone and other rock types, can be recognized at many localities in about the same stratigraphic position.

Vertebrate fossils from the uppermost part of the Galisteo formation have been dated as Duchesnean; one fossil locality lies in beds transitional to the overlying Espinaso volcanics. The Galisteo formation thus in large part postdates the better-known Paleocene and Eocene deposits of the San Juan Basin (Simpson, 1948). However, the Galisteo formation is very thick and probably includes beds which range well down into the Eocene; more adequate paleontologic data are needed to fix the time limits of deposition.

Espinaso Volcanics

The Galisteo formation is overlain conformably by the Espinaso volcanics, comprising water-laid breccia, conglomerate, and tuff, with massive flows and explosive material in subordinate amount. Rock types represented by flows and phenoclasts of the Espinaso volcanics range from andesite to quartz latite, but conspicuously porphyritic latites and quartz latites are by far the most abundant. The rocks are predominantly bluish-, reddish-, and brownish-gray. The bulk of the formation is water-laid; flows are thin and rare at distances of 5 miles from the presumed centers of eruption. Deposition of the Espinaso volcanics succeeded that of the Galisteo formation without break, and the two formations are in many localities separated by a zone of transition. However, the Espinaso volcanics record disruption of the earlier, integrated drainage and burial of the earlier basin of deposition by alluvial fans radiating from centers in the Ortiz Mountains and the Cerrillos Hills. Present thickness of the Espinaso volcanics varies because of mid-Tertiary erosion; a maximum thickness of

1400 feet is preserved in the Espinaso Ridge, north of the Tonque Valley.

Red volcanic conglomerate, gray water-laid tuff, and andesite flows, believed to be part of the Espinaso volcanics, are exposed at several localities along the base of the foothills of the Sangre de Cristo Mountains south of Santa Fe, notably in the Arroyo Hondo (Pl. 1). These beds overlie Precambrian rocks directly and are believed to have been deposited in part of the highland area of Galisteo time.

Intrusive Igneous Rocks

Intrusive igneous rocks of early Tertiary age are exposed in a north-south line of hills from near La Cienega to South Mountain. The intrusives include vents active in Espinaso time and related plutons of diverse form (Stearns, 1952b). Sills and dikes petrogenetically related to the larger intrusions of the line of hills crop out in the Tonque Valley, Galisteo Monocline, and Galisteo Lowland. Intrusive activity in the Galisteo-Tonque area occupies the same interval as does strong deformation of the pre-Abiquiu (?) rocks, described in a later section (Pre-Abiquiu (?) Deformation).

Cieneguilla Limburgite

A series of basalt flows, most of them feldspar-free limburgites, with minor amounts of interbedded basaltic (?) tuff and streamlaid conglomerate, overlies the Espinaso volcanics in the vicinity of La Cienega (Pl. 1). The flows were probably extruded in part from a local vent, represented by a plug underlying Cerro Segura. The Cieneguilla limburgite appears to rest on an erosion surface of considerable irregularity cut in the late Eocene Espinaso volcanics, and is older than the Miocene Abiquiu (?) formation. Presumably equivalent rocks crop out in the canyon of Santa Fe Creek, in the La Bajada escarpment north of Highway 85, and in Huerfano Butte, northeast of Hagan (Pl. 1).

Late Tertiary Rocks

Introduction

The Rio Grande Valley, between Embudo and Socorro, is underlain in large part by deformed alluvial deposits of late Tertiary age. These deposits are concealed over large areas by younger alluvium mantling old erosion surfaces, and by alluvium and slope wash associated with the present cycle of erosion. The name Santa Fe formation (Santa Fe marls) was given by Hayden (1869, p. 167–170) to beds exposed in the vicinity of Santa Fe. The vertebrate fauna found in the type locality is of upper Miocene and lower Pliocene age (Wood *et al.*, 1941, Pl. 1). Recent studies have attributed the bulk of the deformed alluvial deposits in the Rio Grande Valley to the same formation. However, the time equivalence of various beds studied in various areas has not been precisely established, and contained fossils range from late Miocene to late Pliocene in age (Bryan and McCann, 1937; Denny, 1940a; Needham, 1936; Wright, 1946).

The various beds assigned to the Santa Fe formation "conform in their arrangement to a geographic position consistent with the laws of deposition in basins" (Bryan, 1938, p. 205). The work of Bryan and his students has more or less established the thesis that the Santa Fe formation was deposited in a series of basins related to the modern Rio Grande Depression. This thesis can be argued most successfully for late Santa Fe time when the several basins were integrated by a through-flowing river, the ancestral Rio Grande. Much of the Santa Fe formation consists of alluvial fan material, deposited by tributaries to this river. Interbedded lava flows record ancient centers of local volcanic eruption.

Older alluvial deposits, commonly more or less tuffaceous, have been mapped as part of the Santa Fe formation in some areas, and in others separated as distinct formations. These earlier deposits may not be equivalent to each other in time, and the geography of the basins in which they were deposited has not been established. Their present association with the Santa Fe formation in areal distribution suggests a spatial association of the basins in which they were deposited with those of the later Santa Fe. However, the present association may be in part the fortuitous result of post-Santa Fe deformation, which outlined the present Rio Grande Depression.

Abiquiu (?) Formation

In the Galisteo-Tonque area, as elsewhere, the lower portion of the late Tertiary beds possesses persistent differences of color and composition which set it apart from the overlying beds. Exposures are generally gray-white, in contrast with the rather uniform pink-buff of higher beds. The lower beds are generally finer-grained, better sorted, and more cleanly washed. Although the lower beds are predominantly arkosic, a wedge of water-laid tuff is interbedded with them at La Bajada (Pl. 1). This tuff is believed to be a southward extension of the water-laid volcanic débris named Abiquiu tuff (Smith, 1938, p. 944–952). The name Abiquiu is therefore tentatively assigned to the lower portion of the late Tertiary beds of the Galisteo-Tonque area, but as these beds are chiefly nontuffaceous, the name is extended as Abiquiu (?) formation.

East of the Espinaso Ridge, in the valley of the Arroyo Pinovetito, are discontinuous exposures of arkosic sediments, dipping uniformly east. A pace-and-compass traverse across the beds yielded an estimated thickness of 1400 feet. Outcrops are predominantly gray-white, but tan, pink, and orange-buff beds occur locally. Medium- to coarse-grained sand, with some sandy silt, form the bulk of the outcrops. Most beds are only slightly cemented by calcium carbonate. The base of the section contains lenticular, subangular, fine gravels probably reworked from the underlying Espinaso volcanics. Occasional pebbly lenses in the lower few hundred feet of the section carry pebbles of porphyry and chert only 1 or 2 inches across. The porphyries are angular or subangular, and generally thoroughly rotten. The cherts are well-rounded, and were probably reworked from the Galisteo formation. Tinted beds are more common in the upper part of the section. Gravel lenses are more abundant and carry rotten pebbles of porphyry as much as 6 inches across. The upper beds appear to be overlain without break by typical pink-buff alluvial beds of the Santa Fe formation, but the zone of contact is not exposed.

Fine-grained material similar to that in the lower Arroyo Pinovetito section is exposed at scattered localities northward along the base of the Espinaso Ridge. The grain-size decreases, and silty sand predominates in gullies tributary to the Arroyo Vega de los Tanos. Beds of red and green silty clay and white silty marl also occur. Pebble lenses are infrequent above the basal beds. Beds corresponding to the upper part of the Arroyo Pinovetito section may underly the Arroyo Vega de los Tanos, but are not exposed. East of the arroyo are pink-buff beds typical of the overlying Santa Fe formation.

South of the Arroyo Pinovetito, gulches dissecting the Ortiz pediment expose gravelly sand similar to that in the upper part of the Arroyo Pinovetito section. Rotten, subangular pebbles of porphyry 8 inches across were the largest observed. Finer-grained beds such as those in the lower part of the Arroyo Pinovetito section were not observed farther south. These lower beds appear to thin out, either by gradation into coarser beds or under overlap by the higher beds.

Along the edge of the Ortiz pediment north of the Arroyo Vega de los Tanos are numerous outcrops of brownish, non-indurated sands dipping gently southeastward. Lenses of fine gravel containing rotten pebbles of latite porphyry and silt also occur. Contacts with adjacent rocks are not exposed. To the east, the Espinaso volcanics are exposed in gulches dissecting the Ortiz pediment. The volcanics dip westward, and may be separated from the younger beds by a fault. To the west, typical pink and buff beds of the Santa Fe formation are exposed, dipping gently eastward, undoubtedly separated from the older beds by a fault.

An apparently continuous, northward-dipping succession of fine-grained beds is exposed along the La Bajada escarpment between Highway 85 and La Bajada. The beds crop out in a wedge-shaped fault block. To the west, the northward extension of the Rosario fault presumably separates them from typical Santa Fe beds such as are exposed south of Highway 85. To the east, a branch fault separates the block from older Espinaso volcanics. A detailed section was not measured, but the approximate succession of beds from south to north is:

Base of section not exposed	
Massive to porous, crystalline, impure limestone. Gray on fresh surfaces, weathers brown. Rare nodules of chert	15′
Alternating very thin-bedded, impure limestone 6″–2′ thick, and gray or green gritty silt, 1′–4′ thick. The limestones are ripple-marked, and in many places exhibit contemporary deformation (Pl. 2, fig. 1). The silts contain angular fragments of porphyries ¼″–½″ in diameter	125′ (?)
Poorly exposed, but predominantly reddish, clayey silt and gray pebbly silt. Bands of fibrous calcite parallel to bedding of the silt observed *in situ;* calcite rosettes observed in float. Red and green, sticky clay and gray, lime-cemented gravels also occur, but are probably subordinate in amount. Pebbles of latite porphyry in the gravels reach a maximum diameter of 1″	200′ (?)
Gap, concealed by pediment gravels.	100′–200′(?)
Gray-white sand, nearly continuous exposures in a gulch tributary to Santa Fe Creek. Locally stained orange- or pink-buff. No pebbles observed. A single bed of massive, very fine-grained, white tuff, 2′ thick, observed at about two-thirds from the base	250′ (?)
Well-bedded tuff as described below, additional thickness exposed south of Santa Fe Creek	100′ (?)
Gray-white tuff, fine-grained. Thin and very uniform bedding. Forms cliffs at La Bajada (Pl. 2, fig. 2). Occasional thin beds of dark green, sticky clay. Locally stained by wash from buff alluvium underlying the basalt cap of the escarpment, which overlies the tuffs unconformably	275′
	ca. 1000′

Northward along the La Bajada escarpment, landslide débris conceals the beds overlying the tuffs exposed at La Bajada. On the walls of the gulches cutting back into the escarpment at two places are rare exposures of gravelly sand, containing pebbles of gray porphyry. Bedding appears to dip gently north. These outcrops, and pebbles of porphyry in the slope wash, suggest that the tuffs are overlain by several hundred feet of alluvial material similar to, but coarser than, that underlying them. Farther north, andesitic lavas continuous with those of Tetilla Peak are exposed in the escarpment, underlain by brown, fine-grained alluvium containing fragments of red granite. A few exposures of gravelly sand, with pebbles of gray porphyries, occur on the lower portions of the slope. It thus appears that the older lavas of Tetilla and adjacent portions of the Cerros del Rio, which dip gently to the north, form the upper part of a continuous succession including the beds described above. The dis-

PLATE 2. ABIQUIU (?) AND SANTA FE FORMATIONS

FIGURE 1. THIN-BEDDED LIMESTONE OF ABIQUIU (?) FORMATION

North of U. S. Highway No. 85. Note contemporary deformation of beds to right of hammer.

FIGURE 2. WATER-LAID TUFFS OF ABIQUIU (?) FORMATION IN LA BAJADA ESCARPMENT NEAR LA BAJADA

Cliffs are capped by Cuerbio basalt of Mesa Negra; note landslide blocks to left of cliffs. Canyon is that of Santa Fe Creek.

FIGURE 3. RIVER DEPOSITS OF SANTA FE FORMATION

In road-cut along U. S. Highway No. 85, south of Arroyo Tonque. Note sorting and cross-bedding; compare with Fig. 4.

FIGURE 4. COARSE FAN DEPOSITS OF SANTA FE FORMATION NEAR PLACITAS, RESTING UNCONFORMABLY UPON SANDSTONE AND SHALE OF MESAVERDE FORMATION

Man is examining contact.

FIGURE 1

FIGURE 2

FIGURE 3

FIGURE 4

ABIQUIU (?) AND SANTA FE FORMATIONS

FIGURE 1

FIGURE 2

FIGURE 3

PRE-ABIQUIU FAULTS

tinctive brown alluvium underlying the lavas implies that a change in drainage lines within the sedimentary basin occurred at about the time that volcanism began. In White Rock Canyon, these lavas are interbedded with fan deposits and river gravels of the Santa Fe formation (Bryan, 1938, p. 208).

In the Arroyo Hondo, the Espinaso volcanics are overlain by approximately 500 feet of gray-white beds. Loosely consolidated sand predominates, tuffaceous in some beds. Lenses carrying pebbles as much as an inch in diameter are common. In some lenses, the pebbles are of volcanic rocks; in others, of Precambrian crystalline rocks. The fine-grained beds are overlain by pink-buff gravelly sands of the Santa Fe formation.

Between the Arroyo Hondo and Santa Fe, the Abiquiu (?) formation is not exposed. Immediately north of Santa Fe Creek, beds of typical Santa Fe lithology directly overlie the Magdalena limestone, and the Abiquiu (?) formation is absent. Farther north, in the Española Valley, tuffaceous gravel and sand, and at least one basalt flow crop out locally beneath alluvial beds of typical Santa Fe lithology. They lack the red color characteristic of earlier Tertiary sediments and are perhaps equivalent to the Abiquiu (?) formation. First mapped as part of the Picuris tuff (Cabot, 1938, p. 91, Fig. 2), they were later included in the Santa Fe formation (Denny, 1940b, p. 683).

Correlation of these beds with the Abiquiu tuff is based only upon similarity of stratigraphic position and upon the presence of tuffs in the La Bajada escarpment. These tuffs are not strictly comparable to the Abiquiu tuff in the type locality, which is coarser-grained and contains considerable amounts of gravel. However, the Abiquiu tuff, which was deposited by streams entering the Abiquiu quadrangle from the northeast (Smith, 1938, p. 949; Butler, 1946), should be notably finer-grained to the south, as is the tuff at La Bajada. Both the tuff and the underlying, nontuffaceous beds are included in the Abiquiu (?) formation, following Smith's inclusion of thick, basal, granitic gravels in the type locality.

The Abiquiu tuff in the type locality is older than the Santa Fe formation (Smith, 1938, p. 958), dated as upper Miocene and Pliocene (Wood *et al.*, 1941, p. 31, Pl. 1). However, the locations of fossil collections from the Santa Fe formation have never been adequately described. Some may have come in fact from beds of Abiquiu (?) lithology. Butler (1946, p. 104–109) has shown (1) that the Abiquiu tuff is equivalent in part to the Los Pinos formation of northern New Mexico and southern Colorado, and (2) that the Los Pinos formation is probably equivalent to the Fisher latite-andesite of the San Juan Mountains. The Fisher latite-andesite was erupted after the older Creede formation had been eroded to a canyoned topography (Cross and Larsen, 1935, p. 92). The Creede formation has been generally classed as Miocene, following Knowlton's (1923, p. 184) direct correlation of plant remains from the Creede with plant remains from the Florissant lake beds, then regarded as Miocene. However, Gazin (1935) has since suggested that the Florissant may be as old as

PLATE 3. PRE-ABIQUIU FAULTS

FIGURE 1. FAULT ZONE IN WEST LIMB OF TUERTO FAULTED ANTICLINE

Looking north across the Arroyo Tuerto. White beds in middle of picture are Entrada sandstone, overlain by thin Todilto limestone. To left, a strong fault separates these beds from beds of Mancos shale (drab slopes without conspicuous outcrop). To right, outside picture, a strong fault separates Entrada sandstone and Triassic red-beds. Steeply dipping beds in right foreground are Entrada and Todilto, repeated by cross-faults. Ortiz pediment and Ortiz Mountains in background.

FIGURE 2. FAULT BETWEEN TRIASSIC RED-BEDS AND MORRISON (?) SANDSTONE

Looking south in unnamed arroyo tributary to Arroyo San Cristóbal. Cliff in right background is Morrison (?) sandstone capped by thin Dakota (?) formation, south of Arroyo San Cristóbal.

FIGURE 3. FAULT IN MORRISON (?) FORMATION SOUTH OF ARROYO JARA

Looking north. Moderately dipping beds to left are separated from gently dipping beds to right by steep fault of small displacement.

Oligocene (Orellan), and his dating may be extended to the Creede. Hence, the Fisher latite-andesite may conceivably be as old as late Oligocene, although the partly equivalent Abiquiu tuff may conceivably be as young as upper Miocene. It is inferred that the two formations are early or middle Miocene.

The wedge of tuff at La Bajada was carried to the Galisteo-Tonque area by south-flowing streams. The remarkably uniform and relatively thin bedding, however, and the intercalated green clays suggest that the tuffs at La Bajada may have been deposited in standing water. The fine grain and composition of the remainder of the Abiquiu (?) formation suggest erosion of the Galisteo and Espinaso formations in areas without great relief. The common rottenness of the volcanic pebbles may reflect an environment of moderate chemical weathering. The pebbles of Precambrian crystalline rocks in the Arroyo Hondo section could have been entirely reworked from the Galisteo formation, and the fine grain of the containing beds argues against the existence of a highland area with Precambrian outcrop as close as the Sangre de Cristo Mountains. The northward decrease of grain-size in the Tonque Valley suggests drainage lines flowing to the north, northeast, or northwest. The nontuffaceous beds in the La Bajada escarpment, still farther north, are even finer-grained. The thin-bedded limestones undoubtedly accumulated by chemical precipitation in lakes. That the position of the lakes varied from time to time is indicated by interbedded gritty silts. That the lakes occupied the same general area for a considerable period, however, is attested by the thickness of the series as a whole. It is further suggested by the possibility that the tuffs at La Bajada were deposited in a lake.

Santa Fe Formation

General statement.—Beds assigned to the Santa Fe formation in a restricted sense occupy most of the western half of the Galisteo-Tonque area. They also crop out extensively between the Cerros del Rio and Santa Fe, north of Santa Fe Creek. A few exposures were mapped between Santa Fe and the Arroyo Hondo. The formation includes fan deposits, river deposits, travertine, water-laid pumice, and lava flows.

River deposits.—Deep road-cuts along Highway 85 north and south of the Arroyo Tonque expose thick sections of massive, well-sorted gravel and clean-washed sand (Pl. 2, fig. 3). The gravels contain well-rounded and some subangular pebbles averaging 3–6 inches in diameter. The pebbles are chiefly of crystalline and volcanic rocks. Quartzites and fine-grained volcanics, including many types foreign to the Galisteo-Tonque and adjacent areas, predominate. Outcrops are principally gray-white but in places are heavily stained with limonite.

These exposures of gravel and sand stand in marked contrast to those of corresponding texture in the fan deposits. They have been described previously (Bryan, 1938, p. 207; Denny, 1940b, p. 690; Bryan and Upson, unpublished manuscript) as the deposits of a large, through-flowing river, the ancestral Rio Grande. Similar deposits are recorded in the Rio Grande Depression from Buckman, west of Santa Fe (Denny, 1940b, p. 682–683) southward as far as the San Acacia area (Denny, 1940a, p. 89–90). The river deposits are demonstrably equivalent to only the uppermost part of the Santa Fe formation in the Lower Rio Puerco area (Wright, 1946, p. 411–412). It is not known whether the ancestral Rio Grande existed throughout Santa Fe time.

River deposits underlie the spurs between Highway 85 and the Rio Grande north and south of the Arroyo Tonque. Their attitude is in most places indeterminate. East of the highway, they interfinger with fan deposits which have a definite northeasterly dip. Lenses of river gravel are interbedded with typical fan deposits in a zone extending about a mile east of the highway at the Arroyo Tonque and from the highway east to the boundary fault between Bernalillo and Placitas. In a few localities these lenses are cemented by a dirty, black matrix of manganese oxides. In other localities a heavy limonite stain acts as a partial cement. Silt and clay, which are associated with the river gravels in other areas, are rare. They are reported near the pueblo of San Felipe (Bryan and Upson, unpublished manuscript), and crop out in two places south of the Arroyo Tonque.

A bed of water-laid pumice, 15–25 feet thick, crops out east of the highway on both sides of the Arroyo Tonque. Although fragments vary from silt size up to 4-inch pebbles both verti-

cally and on strike, the bed is very persistent. Locally, it is capped by 1–2 feet of porous travertine, and in one locality north of the Tonque travertine occurs both above and below the pumice. Throughout its outcrop, the bed is nearly horizontal, probably dipping gently to the north. It is associated with river gravels in the zone of interbedded river and fan deposits. West of the highway, abundant, discontinuous exposures of white pumiceous gravel and silt in the river deposits probably represent the same horizon. Their discontinuity may be attributed to reworking by the ancestral Rio Grande as it shifted its course from time to time. If they are correctly interpreted as an extension of the same horizon, they confirm the general impression that the river gravels are nearly horizontal.

Fan deposits.—The alluvial fan deposits comprise sandy gravel, gravelly sand, silt, and some silty clay. Individual beds are lenticular, and the texture changes rapidly both laterally and vertically. The proportion of each type varies from exposure to exposure. Most beds are moderately cemented by calcium carbonate. Although variations in color occur, the general aspect of most outcrops is rather uniformly pink-buff.

At Placitas, the Santa Fe formation overlies the Galisteo and Mesaverde formations with angular unconformity (Pl. 2, fig. 4). In this area, the Santa Fe formation consists of coarse, sandy gravel with a minor amount of gravelly sand. The gravel is poorly sorted, although it possesses a rude, discontinuous stratification. Fresh angular and subangular pebbles occur in all sizes up to 2 feet across, and blocks 3–6 feet across are not uncommon. The pebbles include all rock types. In many exposures, gray porphyries are the most abundant rock type. In others, sedimentary rocks are the most abundant. All of the large blocks observed were red sandstone. The gravels are pink to red and are well-cemented with calcium carbonate.

This coarse gravel forms only the lower one or two hundred feet of the formation. Finer-grained material occurs higher in the section at Placitas, northward between the Arroyos Las Huertas and Tonque, and west of the Bernalillo fault. Sandy gravel lenses generally contain pebbles 1–4 inches in diameter, with scattered pebbles 4–8 inches in diameter. A few outcrops of coarser gravel were observed north of Placitas. Gray porphyries, sedimentary rocks, and Precambrian crystalline rocks occur in varying proportions. Between the Arroyos Las Huertas and Tonque, gravel lenses in which about half of the pebbles are porphyries alternate with lenses in which porphyries are absent. Associated with the porphyries are some contact metamorphic rocks which can only have been derived from the Ortiz Mountain-South Mountain chain. Silty sand and sandy silt occur in many outcrops, notably the bluffs north of the Arroyo del Ojito Chico, and presumably form the bulk of the formation. The proportion of finer-grained material increases from east to west (Fig. 4), where the fan deposits interfinger directly with river deposits. The formation in this area is pink, buff, or less commonly gray-white.

The sections (Fig. 4), although they cannot be correlated precisely by recognizable horizons, indicate an aggregate thickness of about 500 feet. Continuous sections of more than a few hundred feet cannot be measured elsewhere, but a continuous section extending 200 feet above the base of the formation was measured at Placitas. Nearly continuous northward dips between this section and the sections measured in the Arroyo Alamos Altos imply that the total thickness of the fan deposits is at least 2000 feet.

Between the Arroyo Vega de los Tanos and the Rosario fault, exposures of the Santa Fe formation are chiefly of pink, sandy gravel cemented with calcium carbonate. Fresh, angular to subangular pebbles in the gravels average 4–8 inches in diameter, but boulders 2 feet across were observed. Most of the pebbles are porphyries derived from the Espinaso volcanics or from the intrusives of the Ortiz Mountains, but sedimentary and crystalline rocks are not uncommon. The latter types are more abundant toward the north and outnumber the porphyries near the northern limit of the area mapped. Silty sand and gravelly sand presumably form a larger part of the formation than the few exposures would indicate. Some of the gravel lenses carry pebbles of rotten gray porphyry, but most of the porphyries are relatively fresh. A bed of fine-grained gray pumice, 15–20 feet thick, crops out near the San Francisco fault (Pl. 1, Loc. 6). The lower half of the bed

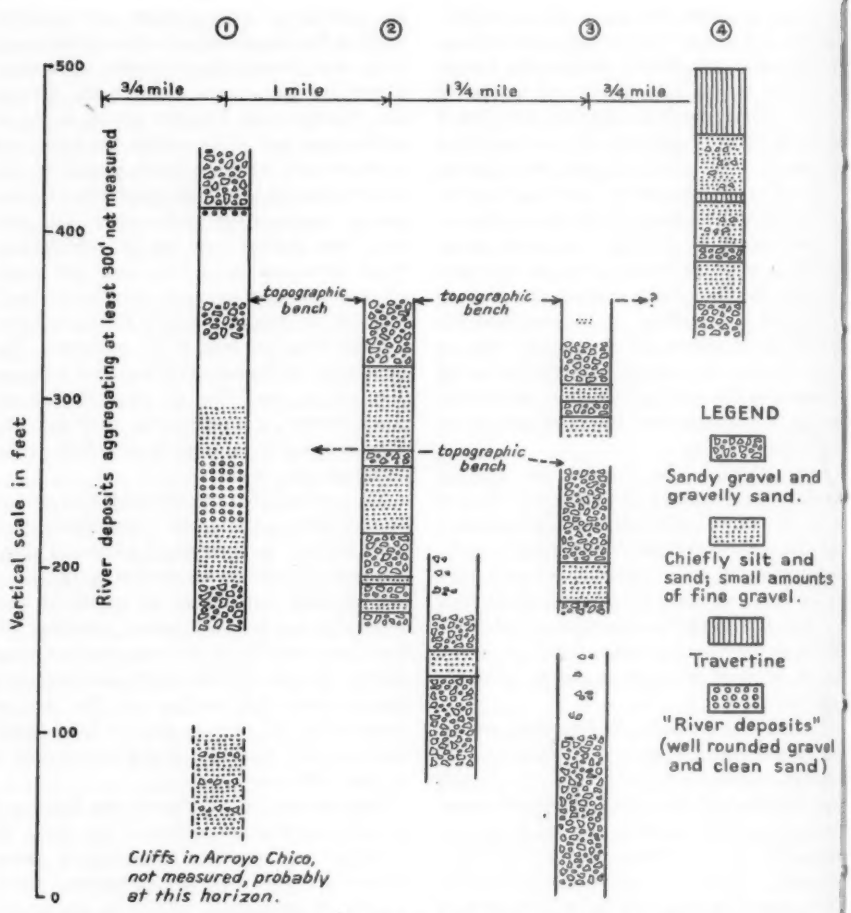

FIGURE 4. SECTIONS OF SANTA FE FORMATION ALONG THE ARROYO ALAMOS ALTOS

is massive and unconsolidated. The upper half is stratified and partially cemented by calcium carbonate.

The Abiquiu (?) formation of the Arroyo Hondo is overlain by several hundred feet of pink and pink-buff arkosic sands and grits. Lenses of gravel contain pebbles of Precambrian granite and metamorphic rocks. Only two exposures of similar material were found between the Arroyo Hondo and Santa Fe. Extensive exposures north of Santa Fe Creek, continuous with those of the Española Valley described by Denny (1940), were not studied in any detail by the writer. Reconnaissance indicates that the fan deposits exposed north of Santa Fe Creek, between Santa Fe and the Cerros de Rio, contain lenses of rather well-rounded gravel and of rather cleanly washed sand. The gravels contain cobbles of quartzite, gneiss and schist ranging up to at least a foot in diameter. The larger cobbles observed were all quartzite. These cobbles are strikingly different from the angular and subangular cobbles carried by the present Santa Fe Creek, and from those of older gravels blanketing the Santa Fe Plateau, which are chiefly reddish granite. Denny (1940b, p. 683) noted that the Santa Fe river gravels at Buckman contain quartzite

cobbles larger than other cobbles. Some of these cobbles were contributed by streams flowing from the Santa Fe area.

Mesa de la Cal.—The Mesa de la Cal at Alamos Altos Ranch, about three-eights of a square mile, is capped by 15–40 feet of light tan, nodular to massive travertine containing scattered, small, irregular cavities. Underlying beds exposed at various places along the scarp of the mesa include highly calcareous silt and silty clay, pink, sandy gravel, and at least one thin bed of travertine. So far as could be determined, these beds are stratigraphically higher than any other beds of the Santa Fe formation exposed in the vicinity. The mesa is for the most part flat, but rises about 50 feet to a small, flat summit in the southeast corner. This irregularity reflects deformation of the capping travertine.

Interbedded lavas.—The Santa Fe formation in the Cerros del Rio consists almost entirely of thick flows of quartz latite, andesite, and andesite-basalt. Basaltic (?) cinders and other ejectamenta are associated with the flows in some localities. The general distribution of Santa Fe lavas (Pl. 1) is taken from an unpublished reconnaissance map loaned by Prof. E. S. Larsen, modified by the writer in a brief reconnaissance using aerial photographs. In White Rock Canyon, flows and intercalated pyroclastics aggregate at least 1200 feet in thickness (Bryan, 1938, p. 208). At the north end of White Rock Canyon, the flows are interbedded with river gravels and fan deposits (Bryan, 1938; Denny, 1940b, p. 683). River gravels crop out in a canyon cut back into the La Bajada escarpment 6 miles north of La Bajada, but their relationship to the lavas is not clear. In the same canyon, and in smaller canyons further south, the lavas are immediately underlain by fine-grained brown alluvium containing small fragments of red granite and basalt.

The Mesa Sant' Ana, west of the Rio Grande at San Felipe, is capped by basaltic flows of at least two ages. The flows are in part interbedded with the Santa Fe formation and in part of Pleistocene age (Bryan, 1938, p. 208). Remnants of two flows, which dip about 10°E, are interbedded with river deposits east of the Rio Grande at San Felipe.

Age and correlation.—No fossils were found in the Santa Fe formation in the Galisteo-Tonque area. Correlation is therefore based upon general continuity of outcrop with adjacent areas. Precise time equivalence of any part of the beds with the fossiliferous beds of the type locality, or of the Rincones de Zia to the west, has not been established. However, it is assumed that general contemporaneity does obtain; fossils from the type locality are regarded as late Miocene and early Pliocene (Wood *et al.*, 1941, Pl. 1).

Norins Well

A deep well, "North Albuquerque Acres No. 2", was drilled by the Norins Realty Company in T11N, R4E, S19, 7 miles north of Albuquerque and about 4 miles west of the Sandia Mountains. The well reached a total depth of 5024 feet and, although it is 8 miles southwest of the Galisteo-Tonque area, is of interest for the thick section of probably Tertiary rocks revealed. A driller's log, furnished by R. L. Bates of the New Mexico Bureau of Mines and Mineral Resources, is given as an appendix.

"Granite wash", "granite boulders", and "coarse gravel" are reported to a depth of 2150 feet, interbedded with "arkose", "sand", and "shale". It seems certain that these materials belong to the Santa Fe formation and date from a time when the Sandia Mountains existed as a highland.

The remainder of the well penetrated beds reported chiefly as "gray sand" and "gray shale". The driller's log indicates no conspicuous break in the sequence, although "brown shale" and "black sand" are interbedded with the predominantly gray materials between 4020 and 4400 feet. No pre-Tertiary sequence known in adjacent areas would be reported as so great a thickness of uniformly gray sand and shale. Cretaceous shales are generally reported largely as brown or black. It seems certain that the Galisteo formation would be reported in appreciable amounts as "red shale" or "red clay". By elimination, it seems necessary to believe that the beds are Tertiary, younger than the Galisteo formation and older than the Santa Fe formation.

Some of the gray beds may belong to the Espinaso volcanics, which are predominantly

gray in the Tonque Valley. The well lies 15 miles southwest of the Tonque Valley, and the grain size of the formation would certainly be reduced in that distance. A thickness in excess of 2874 feet, although at least twice as great as that preserved in the Tonque Valley, is not inconceivable for the Espinaso volcanics. However, such an interpretation would imply that the Norins Well lies in an area not subject to extensive erosion in post-Espinaso, pre-Abiquiu (?) time.

It is also possible that tuffaceous beds of pre-Santa Fe age, other than the Espinaso volcanics, are included in the log. Such beds are known in the western half of the Albuquerque Basin (Bryan and McCann, 1937, p. 817; Wright, 1946, p. 398).

It is also possible that some part of the beds is equivalent to the Abiquiu (?) formation of the Galisteo-Tonque area. However, the Abiquiu (?) formation is not present at Placitas, and the Espinaso volcanics were removed from this area by pre-Santa Fe erosion. This erosion may have contributed to the Abiquiu (?) formation in the Tonque Valley, which contains appreciable amounts of gravel at a greater distance from Placitas. This may be taken to suggest that none of the finer-grained materials of the Norins Well are of Abiquiu (?) age. On the other hand, the well log may be taken as evidence that the Sandia highland was low in Abiquiu (?) time, that fine-grained sediments were deposited at the site of the well, and that the coarser materials of the Tonque Valley came from the southeast.

Assignment of the fine-grained beds either to the Espinaso volcanics or to the Abiquiu (?) formation requires, either because of lack of erosion or because of recorded deposition, that the well lie outside the Sandia "highland" of Abiquiu (?) time. In such a position, both formations might be represented in the well log. If the Sandia highland was high in Abiquiu (?) time, and if Precambrian granites were exposed in the western portion of the area, the Abiquiu (?) formation may be represented by some portion of the upper 2150 feet of coarse, arkosic beds. If the Sandia highland was low, and if Cretaceous and early Tertiary rocks were exposed, the Abiquiu (?) formation probably comprises some portion of the fine-grained gray beds.

It is felt that the latter interpretation is most consistent with the available data. The former interpretation implies that a thick sedimentary cover was stripped from the northern portion of the Sandia Mountains completely in post-Espinaso, pre-Abiquiu (?) time, i.e. that pre-Abiquiu (?) erosion removed substantially more material than has been removed in Abiquiu (?) and subsequent time. It further implies that the greater part of the uplift of the Sandia Highland relative to the Albuquerque Basin was accomplished in pre-Abiquiu (?) time. In view of the known magnitude of intra-Santa Fe and post-Santa Fe erosion in adjacent areas, and of the known magnitude of post-Santa Fe faulting, either of these implications is unlikely.

It is therefore suggested that the 2150 feet of coarse-grained beds in the Norins Well belong to the Santa Fe formation in a restricted sense. The remaining, fine-grained beds probably include equivalents of both Espinaso volcanics and Abiquiu (?) formation, but the respective thickness of either formation cannot be safely estimated.

Quaternary Rocks

The Quaternary history of the Galisteo-Tonque area has been dominated by erosion. Nevertheless, erosion has not been everywhere continuous, and there are extensive Quaternary formations (Pl. 1) which overlie pre-Quaternary rocks with strong angular unconformity. They comprise:

(1) thick gravel mantling the high surface of the Estancia Valley, possibly as old as the Santa Fe formation, but mapped together with

(2) the Tuerto gravel, 5 to more than 150 (300?) feet thick, a record of aggradation in response to deformation during the final stages of development of post-Santa Fe Ortiz pediment;

(3) basalt flows and associated cinder cones, extruded in and near the Cerros del Rio and interbedded with the Tuerto gravel, tentatively correlated with the Cuerbío basalt (Bryan, 1938, p. 218);

(4) pediment alluviums associated with three erosion surfaces younger than the Ortiz pediment, not shown in Plate 1;

TABLE 1.—TERTIARY STRUCTURAL HISTORY OF THE GALISTEO-TONQUE AREA

Time	Stratigraphy	Diastrophism
Post-Santa Fe	Erosion; Tuerto gravel and Cuerbío basalt	Bernalillo, Buena Vista, San Francisco, and Rosario faults; uplift of Sangre de Cristo Mts. and warping of blocks between major faults (Post-Santa Fe faulting; Pl. I, Fig. 9)
Pliocene	Santa Fe formation	Differential subsidence of sedimentary basin (Santa Fe basin of deposition; Fig. 8)
Miocene	Abiquiu (?) formation	Differential subsidence of sedimentary basin (Abiquiu basin of deposition; Fig. 7)
—?—?—?	Cieneguilla limburgite	Quiescence?
Oligocene	Erosion	Tijeras, Los Angeles, Las Jollas fault systems; Sandia, Tuerto, La Madera faulted anticlines; Uña de Gato and Galisteo basins; fault boundary of Glorieta Mesa; Minor faults and folds in Tonque Valley (Pre-Abiquiu (?) deformation; Pl. 4)
Eocene	Espinaso volcanics	Intrusion of stocks in Cerrillos Hills, Ortiz Mountains, San Pedro and South Mountains; dikes and sills in Tonque Valley and Galisteo Lowland (Intrusive igneous rocks; Pl. I)
	Galisteo formation	Differential subsidence of sedimentary basin (See Stearns, 1943, p. 308–309, Fig. 6)
	Erosion	Gentle folding (See Stearns, 1952a, Fig. 1)
Paleocene		
K	Mesaverde formation	Not discussed

(5) post-Cuerbío basalt in the Mesa Sant' Ana;

(6) late Pleistocene and Recent alluvium in the present valley floors, of which only restricted areas are shown in Plate 1.

Discussion of these formations, and of the long period of erosion during which they were formed, is deferred to a subsequent paper.

STRUCTURE

Introduction

The pre-Tertiary structural history of the Galisteo-Tonque area has been summarized by Read *et al.* (1944). The Tertiary structural history is summarized in Table 1. Mild early Tertiary deformation has been discussed in a previous paper (Stearns, 1943), and deformation contemporary to deposition of the Abiquiu (?) and Santa Fe formations is discussed later (Late Tertiary History of North-central New Mexico). These episodes of deformation, inferred from stratigraphic evidence, are revealed in Plate 1 by variations in thickness of individual formations. The major structural features shown in Plate 1 were formed during two intervals of more intense deformation, pre-

Abiquiu (?) and post-Santa Fe. The younger, post-Santa Fe structures comprise strong, gravity faults arranged en échelon, with adjustment by warping of the intervening blocks. The earlier structures comprise a series of complexly faulted folds, developed in a block northwest of a prominent, northeast-trending fracture system.

Post-Santa Fe Faulting

Introduction.—The Galisteo-Tonque area lies on the east side of the Rio Grande Depression, in the area of offset between the Albuquerque-Belen Basin and the Española Basin (Fig. 6; Bryan, 1938, p. 209–211). The present structural pattern of the Galisteo-Tonque area (Fig. 7; Pl. 1) is therefore dominated by normal faults associated with the border of the Rio Grande Depression, principally of post-Santa Fe age. The east border of the Albuquerque sub-basin is a great fault at the west base of the Sandia Mountains. The east border of the Española Basin is a faulted unconformity, at the west base of the Sangre de Cristo Mountains (Denny, 1940b, p. 685–686). The off-set between the two fault borders is marked by a series of faults en échelon, downthrown to the west.

The north end of the Sandia Mountains lies within the Galisteo-Tonque area. The Bernalillo fault (Pl. 1) is the northernmost member of a fault system at the west base of the Sandia Mountains. The first major fault to the east, the San Francisco fault, separates the older rocks of the Tonque Valley from the Santa Fe formation in the Rio Grande Depression. An area of complex structure at Placitas occurs in the area of offset and adjustment between these two faults and determines the north end of the Sandia block. A second major fault, the Rosario fault, extends northward from the west base of the Ortiz Mountains. Adjustment between the San Francisco and Rosario faults is recorded by northeastward dips in the Santa Fe formation, changing to northerly dips in a synclinal trough at the southern termination of the Rosario fault. The faulting characteristic of the east border of the Española sub-basin has not been recognized south of the city of Santa Fe. Instead, the mountain border south of Santa Fe appears to be monoclinal, and the Tertiary rocks dip gently westward above unconformable contacts. The area between the mountain front and the Rosario fault is broadly synclinal, as pre-Santa Fe rocks are exposed both east and west of the Santa Fe Plateau. However, structural details are concealed by extensive Quaternary gravels and basalt flows.

Major en échelon faults.—The Bernalillo, San Francisco, and Rosario faults are fault zones, 50–300 feet wide, commonly marked by a jumbled mass of rocks from all parts of the stratigraphic column and generally poorly exposed. Springs are common. Individual fault planes which are exposed are steeply-dipping normal faults. Although irregular in detail, the traces of the major fault zones appear to be essentially straight over considerable distances, and are so shown in Plate 1. The fault zones are certainly steep and presumably normal.

The strikes of many of the pre-Santa Fe rocks swing sharply as they approach the fault zones, in anticlinal structures. This phenomenon is believed to be large-scale drag folding; it may indicate that the faults were initiated as monoclines, at least at depth. Adjacent to the fault zones, the pre-Santa Fe rocks are generally complexly fractured. Triassic, upper Cretaceous, and early Tertiary rocks near the San Francisco fault zone are broken into a giant fault breccia, which preserves only a suggestion of stratigraphic continuity for a distance of 100–300 feet before passing over into the fault zone proper. Deformation of the Santa Fe formation is not as readily observed. Scattered exposures, however, yield dips and strikes discordant with the general attitude of the Santa Fe formation, and some deformation does exist. In the Mesa de la Cal (Pl. 1, Loc. 4), a small anticline in the Santa Fe formation adjoins the San Francisco fault zone. Neither the Santa Fe formation nor the adjacent Paleozoic rocks at this locality are as intensely brecciated as are the Mesozoic rocks further north.

On those faults which separate the Santa Fe formation from older rocks, the stratigraphic displacement at any locality is the sum of (1) the local thickness of the Santa Fe formation and (2) the thickness of rocks originally intervening between the rocks now exposed adjacent to the fault and the rocks locally underlying the pre-Santa Fe unconformity. In the present lack of data on either the thickness of the Santa Fe formation or the distribution of older rocks

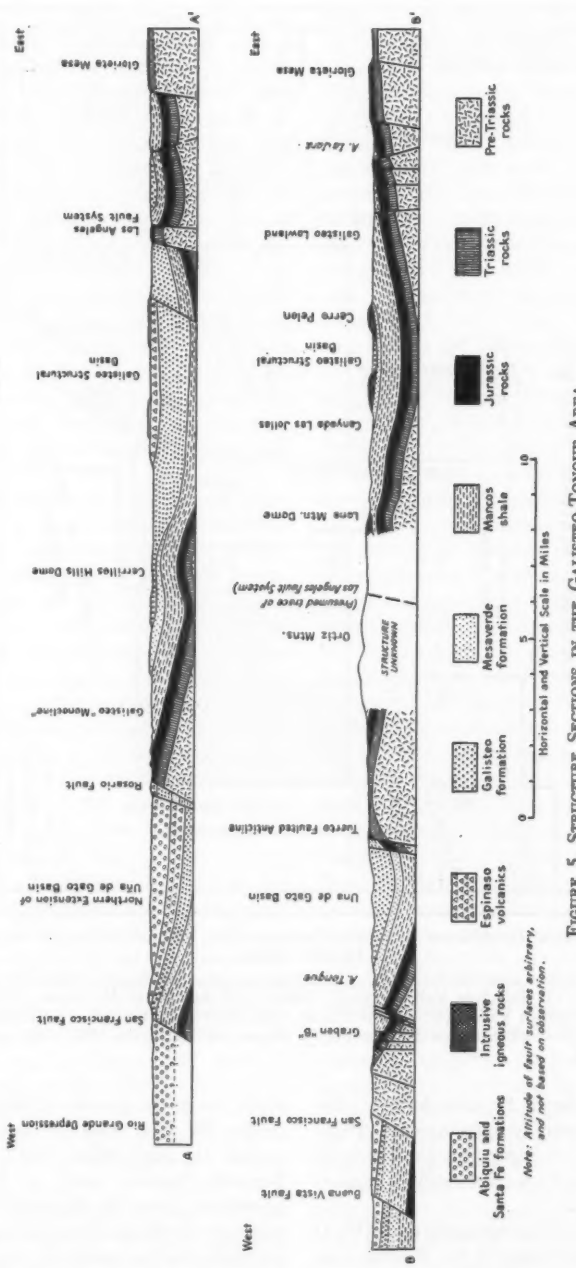

FIGURE 5. STRUCTURE SECTIONS IN THE GALISTEO-TONQUE AREA

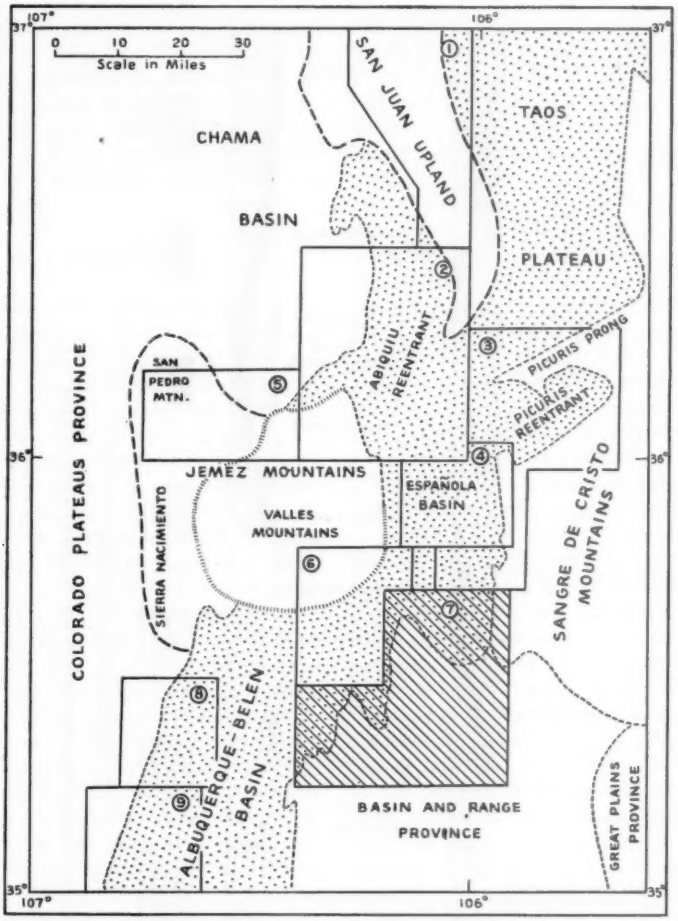

FIGURE 6. RIO GRANDE DEPRESSION IN NORTH-CENTRAL NEW MEXICO (STIPPLED) SHOWING AREAS OF RECENT STUDY

(1) Tusas-Tres Piedras area (Butler, 1946); (2) Abiquiu quadrangle (Smith, 1938); (3) Picuris-Santa Fe area (Cabot, 1938), (4) Española Valley (Denny, 1940b); (5) San Pedro Mountain (Church and Hack, 1939); (6) Santo Domingo Valley (Bryan and Upson, unpublished manuscript); (7) Galisteo-Tonque area of present report; (8) Northern Ceja del Rio Puerco (Bryan and McCann, 1937); (9) Lower Rio Puerco area (Wright, 1946).

beneath the pre-Santa Fe unconformity, displacements are generally indeterminate. Structure sections (Fig. 5) imply that displacements are large, and data from a few localities support this hypothesis.

Bernalillo fault.—The Bernalillo fault (Pl. 1) separates the older rocks of the Placitas area from the Santa Fe formation to the west. It is named from springs (Ojito de Bernalillo) which rise in the Arroyo de Bernalillo where it crosses the fault zone. At and south of the springs, the fault strikes N10°E. South of the Bernalillo-Placitas road, a breccia formed principally from the Morrison (?) formation separates the Santa Fe formation from Triassic red beds. Farther south, the fault zone is concealed by pediment gravel, but presumably continues with increasing displacement. The

zone must be offset to the west around the foothills of the Sandia Mountains, which have a linear front striking N45°E, and is probably continuous with the strong fault or fault zone on the east side of the Albuquerque-Belen Basin. Northeast of the springs, the Bernalillo fault strikes N45°E and enters the Santa Fe formation, in which the fault can be traced only half a mile. The Bernalillo fault is, at least in part, of post-Santa Fe age. Evidence of pre-Santa Fe movements is lacking.

San Francisco fault.—This fault zone, named from springs which rise in it at San Francisco, separates the older rocks of the Tonque Valley from the Santa Fe formation to the west. Its general northeasterly strike is accomplished by short sections most of which strike alternately N10°E and N45°E. Stratigraphic displacement is about 5000 feet at Tecolote (Loc. 5, Pl. 1, northeast of Placitas; Sec. B-B′, Fig. 5) where a jumble of Triassic and Jurassic rocks separates the Mesaverde formation from the Madera limestone. At the Arroyo Vega de los Tanos, stratigraphic displacement is the thickness of the Santa Fe formation preserved west of the fault, which is probably near the maximum thickness preserved in the area (2000 feet?). Bryan and Upson (unpublished manuscript) have shown that the fault probably dies out to the north within the Santa Fe formation. The fault is in part post-Santa Fe, but it has been shown in an earlier section (Santa Fe formation) that the fault was probably active during late Santa Fe time.

Rosario fault.—The Rosario fault was named by Bryan and Upson (unpublished manuscript) from Rosario Siding (Pl. 1), near which it crosses Galisteo Creek. The fault zone separates Mesozoic rocks in the Galisteo Monocline from the Santa Fe formation in the Santo Domingo Valley. The Rosario fault proper can be traced 6 miles south from Rosario Siding before it passes underneath Quaternary gravel west of the Ortiz Mountains. It presumably continues for some distance south of the exposures mapped, but must die out southward. Nearly continuous exposures in the Arroyo Tuerto show clearly that the Rosario fault does not extend that far, and the available exposures north of the Arroyo Tuerto can be grouped in an anticlinal structure without the assumption of a fault (Fig. 5, Sec. AA′). A short distance south of Galisteo Creek, a branch fault cuts southwest from the Rosario fault, interposing a wedge of Espinaso volcanics and Abiquiu (?) formation between the Mesozoic rocks and the Santa Fe formation.

North of Rosario Siding, the Rosario fault can be traced as far as U. S. Highway 85. Exposures of the Santa Fe formation west of the fault were not found north of the highway, but exposures of pre-Santa Fe rocks to the east in the associated La Bajada escarpment imply that displacement continues several miles farther north. West of Tetilla Peak, the escarpment swings northwestward across the projection of the fault trace to the Rio Grande with an irregular pattern. This may be taken to suggest that the Rosario fault dies out west of Tetilla Peak, but detailed mapping of the lavas of the Cerros del Rio will be necessary to show their structural relationships.

U. S. Highway 85 crosses the La Bajada escarpment in an irregular re-entrant eroded in rocks broken by a complex fault system at the junction of two strong branch faults with the Rosario fault. The branch faults, striking N25°E, are downthrown to the northwest, so that a wedge of the Espinaso volcanics and Abiquiu (?) formation is interposed between the Santa Fe formation and the Mesozoic rocks. The branch faults are exposed in the canyon of Santa Fe Creek and must continue some distance farther north. That their continuity may be considerable is suggested by two lines of Quaternary cinder cones northeast of Tetilla Peak, the strikes of which lie on projections of the faults exposed farther south. The alignment of the cinder cones implies genetic association with linear fractures, and continuation of the faults would provide the necessary fractures.

Bryan and Upson have shown, from the lithology of the Santa Fe formation in the Santo Domingo Valley, that the Rosario fault had no physiographic expression in Santa Fe time. There is no evidence that the Rosario fault and the associated branch faults were active in other than post-Santa Fe time.

Placitas area.—The Sandia Mountains have been carved from a great fault block, raised with respect to the Santa Fe formation in the Rio Grande Valley along a fault zone of which the total displacement southwest of the Galisteo-Tonque area (at the Norins Well, described in

an earlier section) may be in excess of 15,000 feet. As indicated above, the Bernalillo fault is believed to be the north end of this fault. A complex structure at Placitas marks the north end of the Sandia Mountains and provides adjustment in the area of offset between the Bernalillo and San Francisco faults. Paleozoic and Mesozoic rocks dip northward off the mountain mass and underneath the Santa Fe formation. Their structure is largely concealed by gravel on pediment remnants, alluvium in valley bottoms, and intervening slope wash. General features are presented in Pl. 1, but more detailed study would undoubtedly modify the map.

The Placitas area is divided by the north-south Buena Vista fault zone. West of the fault, the Paleozoic-Mesozoic sequence dips northward at angles of 38° to 60°; smaller dips were recorded in a few localities. No faults of large displacement break the sequence.[1] East of the north-south fault, the northward-dipping sedimentary sequence is faulted down against the Sandia mass along a system of cross-faults with an average total displacement of 2500 feet. This system joins the San Francisco fault to the east. The area mapped as Dockum Group is largely concealed, and is undoubtedly affected by cross-folding or faulting. The remainder of the sequence dips northeast at angles of 20° to 50°.

Relative uplift of the Placitas area in pre-Abiquiu (?) time is indicated by the absence of the Abiquiu (?) formation in that area. Some of the complex structure at Placitas may date from this earlier period of deformation. However, structures belonging to separate intervals of deformation cannot be differentiated, and most of the structures mapped at Placitas are probably of post-Santa Fe age.

Tonque Valley and area to the north.—Adjustment between the San Francisco and Rosario fault systems is reflected by general northeasterly dips in the Abiquiu (?) and Santa Fe formations east of the Espinaso Ridge. To the north, northeasterly dips prevail and the block is tilted obliquely to the two fault systems (Fig. 5, Sec. AA'). East of Huerfano Butte (Pl. 1), the Abiquiu (?) formation strikes east-west, so that the structure is synclinal near the southern termination of the Rosario fault. The pre-Abiquiu (?) rocks in the Tonque Valley conform to the synclinal pattern of deformation, and presumably owe much of their present dip to post-Santa Fe deformation.

Santa Fe plateau.—The area between Mesa Negra and the Sangre de Cristo Mountains is blanketed by Quaternary gravel, and the underlying structure can only be surmised. Westward-dipping beds of the Santa Fe formation exposed in the Arroyo Hondo must underlie much of the plateau. Older beds cannot be distinguished from Quaternary gravels in well logs from the northern part of the plateau, some of which are more than 400 feet deep; the logs may be taken to suggest either that the Santa Fe formation is present, or that the Quaternary gravels are thick. In the Arroyo Hondo, the contact between Tertiary and Precambrian rocks is sedimentary, and the Sangre de Cristo Mountains south of Santa Fe appear to be bounded by a tilted unconformity, unbroken by faulting. The western boundary of the Santa Fe formation in the plateau is nowhere exposed. North of Santa Fe Creek, beds of the Santa Fe formation dip westward at Agua Fria and are essentially horizontal at the eastern boundary of the Cerros del Rio. The pre-Quaternary rocks of the Santa Fe Plateau therefore appear to be disposed in a broad syncline plunging northwestward toward the Cerros del Rio. Farther north, in the Española Valley, general westerly dips prevail (Denny, 1940b, Pl. 1). A structural boundary between this area and the Santa Fe Plateau cannot be established.

Pre-Abiquiu (?) Deformation

Introduction.—Structures younger than the Espinaso volcanics but older than the Abiquiu (?) formation are exposed in the Galisteo Lowland and in the Tonque Valley, separated by the Ortiz Mountains, never mapped in detail. Relationships between the two areas are not apparent in Plate 1, so the pre-Abiquiu (?) structures are shown in Figure 7, which in-

[1] Read *et al.* (1944) show several east-west faults within the Magdalena group west of the Buena Vista fault. The writer, less familiar with the subdivisions of the Magdalena group established by those workers, had interpreted the complex relationships observed as the combined effects of irregularities in the pre-Magdalena and modern topographies. It is presumed that faulting does exist as they show it.

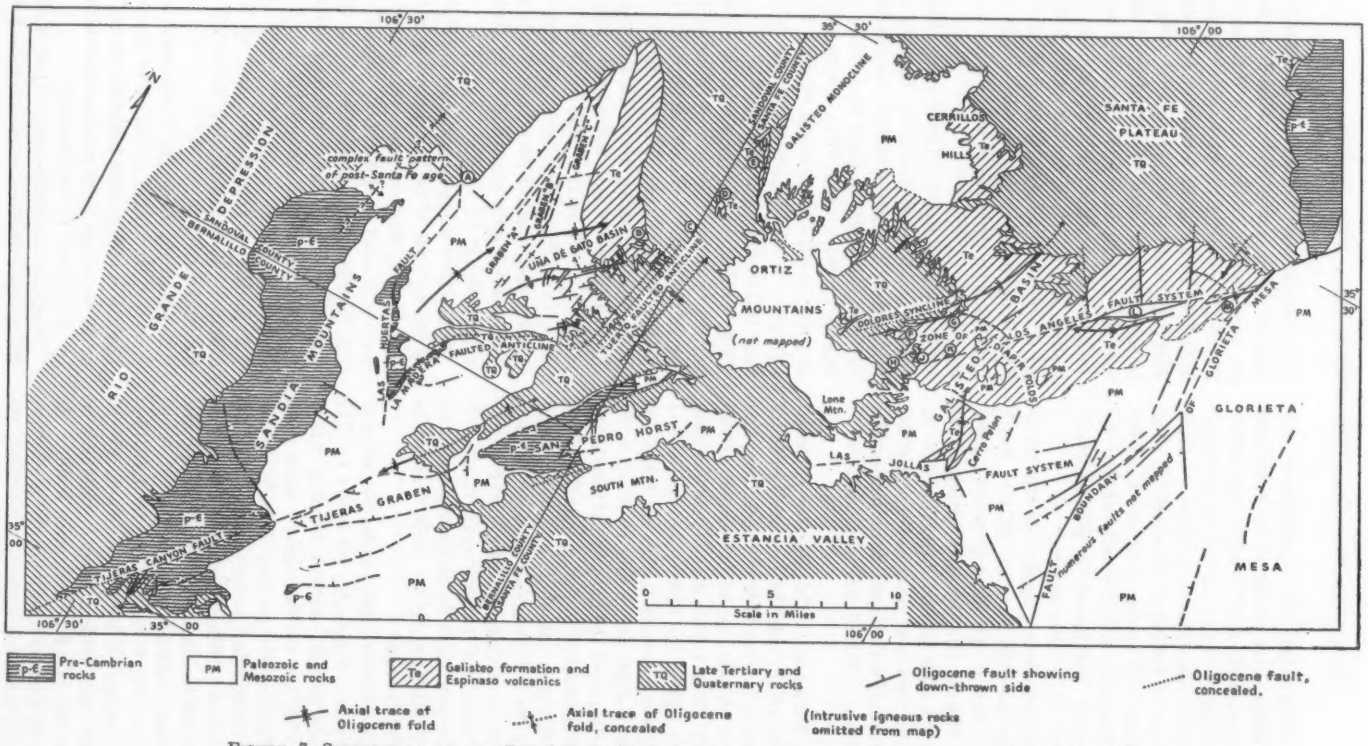

FIGURE 7. STRUCTURES OF THE PRE-ABIQUIU(?) ROCKS IN THE GALISTEO-TONQUE AND ADJACENT AREAS

cludes a considerable area southwest of the Galisteo-Tonque area (Data for the latter area are taken from Read *et al.*, 1944, geologic map). Contemporaneity cannot be demonstrated for all of the structures shown; available evidence for inclusion of individual structures is summarized below. All appear to fit into a reasonable areal pattern, and it is believed that the entire pattern was formed in post-Espinaso, pre-Abiquiu (?) (Oligocene) time.

Igneous rocks were emplaced in the north-south line from the Cerrillos Hills to South Mountain during the interval post-Espinaso and pre-Abiquiu (?). Intrusive activity thus falls into the same general interval of time as do the structures next described. Local structures associated with intrusions in the Cerrillos Hills (Stearns, 1952b) are not included in Plate 4.

Sandia Mountains.—The Abiquiu (?) formation is not present near Placitas, and the Santa Fe formation overlies beds of the Galisteo and Mesaverde formations with angular unconformity (Pl. 1). It follows that an area including the north end of the Sandia Mountains and extending some distance farther north was uplifted and subject to erosion in Abiquiu (?) time. Furthermore, the fine grain of beds in the Norins Well assigned to the Abiquiu (?) formation implies that the uplift and initial rapid erosion in the Placitas area were largely accomplished in pre-Abiquiu (?) time. However, the present structural pattern near Placitas is principally post-Santa Fe. Structures associated with the pre-Abiquiu (?) uplift cannot be specifically identified.

The Las Huertas fault, on the east slope of the mountains, is probably pre-Santa Fe. The fault trace includes sections trending N10°E and N25°W, and can be traced 7 miles south of the Galisteo-Tonque area (Fig. 7). Stratigraphic displacement varies from less than 100 to more than 1500 feet, down to the west. One mile east of Placitas, a parallel fault of opposite displacement outlines a small graben. At Tecolote (Loc. A, Fig. 7; Loc. 5, Pl. 1) the Las Huertas fault is cut off by the post-Santa Fe San Francisco fault system. Farther southwest, at Placitas, the Santa Fe formation overlies the Mesaverde formation with angular unconformity (Pl. 1). Both formations dip northeast, the older formation more steeply. If these attitudes were maintained without interruption, the Galisteo formation should underlie the Santa Fe formation at Tecolote, as it does north of Placitas. Instead, the Mesaverde formation underlies the Santa Fe formation at Tecolote. This repetition of the Mesaverde formation beneath the pre-Santa Fe unconformity shows that the Las Huertas fault continues beyond its intersection with the San Francisco fault system, and that it is pre-Santa Fe, probably pre-Abiquiu (?), in age. Its effect in mid-Tertiary time, as now, was to raise a subsidiary fault block on the east side of the larger Sandia uplift.

Uña de Gato Basin.—The Tonque Valley is eroded in Mesozoic and early Tertiary rocks disposed in a broad syncline plunging northeastward, the Uña de Gato Basin (Fig. 7) of Darton (1928a, p. 100). The east limb of the syncline is cut off by faults (Tuerto faulted anticline). In the broad heel of the syncline, near Hagan, beds dip from 20° to 50° northward, crenulated by gentle folds plunging northeastward and broken by complex faulting; the folds and faults do not appear to affect the late Tertiary rocks. Beds in the west limb of the syncline strike nearly north-south and dip from 10° to 30° east, in essential conformity to beds of the Abiquiu (?) and Santa Fe formations. Thus, the general synclinal structure is the resultant effect of post-Santa Fe deformation imposed upon beds which were deformed in pre-Abiquiu (?) time.

Gentle folds plunging northeast are indicated by the attitude of beds and outcrop patterns near Hagan. Axes are not well defined, but may be located on Plate 1 as follows:

1. Synclinal, from exposures in the Arroyo Tuerto, 3 miles southeast of Hagan, southwest to exposures in the Arroyo Tonque, 3 miles south of Hagan.

2. Anticlinal, 1 mile southeast of Hagan, obscured by faults roughly coincident with the axial trace.

3. Synclinal, from the Cretaceous beds 2 miles west of Hagan northeast to the area of Espinaso outcrop near Huerfano Butte. Dips in the west limb of the syncline were intensified by post-Santa Fe deformation.

Northeast of Huerfano Butte (Loc. B, Fig. 7), the Abiquiu (?) formation maintains a northwest strike and gentle northeasterly dips as

far east as exposures exist. It thus appears to bevel the northeast-trending folds in the older beds to the south.

Fault surfaces are generally not exposed in the Tonque Valley, but exposures on both sides of a fault define its position closely in many localities. Most of the faults were located by discontinuities of prominent horizon-markers, but many can also be traced by stratigraphic discordance of less conspicuous intervening beds. From available data, individual faults are straight lines at a scale of 2 inches to the mile. However, the fault surfaces are presumably somewhat wavy and irregular in detail. In the irregular local topography, the regularity of trace of the faults implies that they are nearly vertical. Approximate measurements on fault surfaces of small displacement, exposed on steep slopes, yielded dips of 65°, 80°, and 85°. The throws of each are in the direction of dip, and the faults are normal. In one fault zone crossing the Dakota (?) hogback (included in lower Mancos shale, Pl. 1) east of the Arroyo Tonque, three fault surfaces dip 85° east. One of the three has reverse displacement, but displacement on the other two and in the zone as a whole is normal. In a fault zone between Morrison (?) sandstone and Mancos shale, trending NNW from Tonque (Pl. 1), isolated fault surfaces dip from 45°E to 45°W. However, these surfaces separate individual blocks in what is essentially a giant fault breccia. Many prominent outcrops of silicified sandstone are separated by vertical faces from adjacent small outcrops of shale, and the main fault surfaces enclosing the giant breccia must be essentially vertical.

Almost all of the known faults occur in pre-Tertiary rocks; only one was shown to offset the base of the Galisteo formation. Faults of small displacement could occur unrecognized within the Galisteo formation, but faults of large displacement would certainly have been recognized from available outcrops. Restriction of faulting to the pre-Tertiary rocks must arise from greater competency of the Galisteo formation, for the faults were formed in post-Espinaso time.

Individual faults trend N10°W, N10°E, and N45°E, with few exceptions, and are interpreted as gravity faults. Many faults are terminated by cross-faults of the same age; in no case do offset relations indicate diverse age. As mapped, the faults fall into two principal groups. Several faults of small displacement, southeast of Hagan, coincide approximately with the crest of an anticline in Cretaceous beds. A thick sill is conspicuously offset by several faults, but porphyry dikes radial to the Ortiz Mountains cross faults in three localities without visible offset. Several faults on the west side of the Tonque Valley outline a series of three grabens (grabens A, B, and C; Fig. 7). The southernmost of the grabens (graben A, Fig. 7) trends approximately N10°E and appears to continue south of the area mapped; the other two (grabens B and C, Fig. 7) trend approximately N10°W. As indicated in section AA′ (Fig. 5), the grabens interrupt but do not themselves contribute to easterly dips in the Uña de Gato Basin. Again, thick sills at two different horizons in the Mancos shale are broken by faults on the east side of graben A (1½ miles west of Hagan, Pl. 1), yet a radial basalt dike crosses graben B (just north of previous location) without visible offset.

Tuerto faulted anticline.—Three miles east of Hagan, the east limb of the Uña de Gato Basin is broken by a strong fault zone exposed in the Arroyos Chinana and Tuerto (Pl. 1; Pl. 3, fig. 1). The fault zone includes two major faults striking N10°E and numerous cross-faults of small displacement. The combined stratigraphic displacement in the fault zone is about 3000 feet in the Arroyo Chinana and about 2000 feet in the Arroyo Tuerto. Furthermore, in the Arroyo Tuerto, discontinuous exposures of Triassic beds define an anticlinal axis, plunging northward, about a mile east of the fault zone. As shown in section BB′ (Fig. 5), the few scattered outcrops of pre-Tertiary rocks on the pediment surface north of the Arroyo Tuerto imply that the axial trace parallels the fault zone in a general way, plunging northward. The fault zone thus occurs in a larger zone by which beds are displaced down to the west, in part by flexure and in part by faulting. The entire structure is named the Tuerto faulted anticline (Fig. 7), from the arroyo where it is most completely exposed.As shown in section BB′ (Fig. 5), the structure might also be termed a faulted monocline, forming the west boundary of a block uplifted and tilted eastward. Northeast of Huerfano Butte,

in the gulch of the Arroyo de los Angeles (Loc. C., Fig. 7), beds of the Abiquiu (?) formation are poorly exposed across a projection of the Tuerto faulted anticline. The beds dip east more strongly than do those farther west and were probably deformed by renewed movements in the fault zone. However, Cretaceous rocks crop out near by, and the Abiquiu (?) formation therefore bevels the anticlinal structure and rests on pre-Tertiary rocks; at least 5000 feet of beds were removed by pre-Abiquiu (?) erosion. Still farther north (Loc. D and E, Fig. 7), Galisteo formation and Espinaso volcanics crop out east of a projection of the anticlinal axis. They were not eroded in pre-Abiquiu (?) time. This northward decrease in magnitude of pre-Abiquiu (?) erosion implies that the magnitude of the Tuerto faulted anticline decreased northward.

These relationships, although by no means as clearly exposed as one could wish, show that the Tuerto faulted anticline was formed and eroded in pre-Abiquiu (?) time. Renewed movements probably occurred in the fault zone in post-Abiquiu (?) time; whether or not these movements were also post-Santa Fe cannot be demonstrated.

La Madera faulted anticline.—The rocks of the Uña de Gato Basin dip off a northeast-trending arch south of the Galisteo-Tonque area (La Madera faulted anticline, Fig. 7). The west end of the arch is broken by a branch of the Las Huertas fault; it is probable, as inferred by Read *et al.* (1944, geologic map), that the fault is continuous, with changing strike, northeast to the Arroyo Tonque, although concealed by Quaternary gravels. The inferred trace of the fault is approximately coincident with the crest of the arch, which is thus a faulted anticline. East of the Arroyo Tonque and south of the Arroyo Tuerto, the Tuerto faulted anticline and the La Madera faulted anticline converge in an area concealed by Quaternary gravels. The two structures must form an essentially continuous boundary to the Uña de Gato Basin.

Tijeras fault system.—The Arroyo Tuerto, for a distance of a mile east of state highway 10 (Pl. 1), follows a fault between Mancos shale to the north and Permian and Triassic rocks to the south. Farther east, the fault is concealed, but its displacement must continue between the Ortiz Mountains, in which early Tertiary igneous rocks have invaded Cretaceous sedimentary rocks, and San Pedro Mountain, in which early Tertiary igneous rocks have invaded rocks of the Magdalena group. The fault has been mapped for 8 miles south of the Arroyo Tuerto (Read *et al.*, 1944; Fig. 7) as a northeast-trending, steep thrust fault.

Although this fault is poorly exposed in the Arroyo Tuerto, it belongs to a fault system of great regional importance. Both Darton (1928b) and Read *et al.* (1944), although their maps differ in detail, show the fault system as continuous southwest of the Galisteo-Tonque area for 26 miles, to the west base of the Manzano Mountains. Immediately south of the Galisteo-Tonque area, the steep thrust fault separates a sharp synclinal structure in Permian and Triassic rocks to the northwest from Precambrian and gently dipping Pennsylvanian rocks to the southeast. The Pennsylvanian rocks of San Pedro Mountain locally overlie unconformably and locally are down-faulted against Precambrian rocks; farther southwest, the area of Precambrian outcrop is much greater. These rocks, in turn, are separated from the Permian rocks of South Mountain by a parallel fault or faulted monocline, of opposite displacement. San Pedro Mountain and the broad area of Precambrian outcrop southwest of it thus have the general structure of a horst (San Pedro horst, Fig. 7). Still farther southwest, the Cretaceous rocks of the Tijeras coal field crop out in a synclinal wedge bounded on both sides by faults. The graben thus outlined (Tijeras graben, Fig. 7), and the faults which bound it, continue the trend of the San Pedro horst and its boundary faults. A sharp transverse monocline separates horst and graben. At the southwest end of the Tijeras graben, the bounding faults converge and are continued by a single fault (Tijeras Canyon fault, Fig. 7), with displacement down to the southeast, which separates the Sandia and Manzano Mountains structurally and determines the position of the topographic gap between them (Reiche, 1949, p. 1203). The southwesternmost portion of the Tijeras Canyon fault experienced some movement in late Tertiary time.

This long, northeast-trending fault system is here called the Tijeras fault system. It is double through much of its distance and outlines a

linear slice in which deformation has been somewhat independent of deformation in adjacent areas. The present position of rocks exposed within the slice relative to the rocks of adjacent areas varies on strike; in part, the slice is a graben, in part, a horst. Individual members of the fault system are apparently steep and chiefly normal, but one fault has been mapped as a steep thrust. The southwest end of the system, a single fault, may extend beyond its known outcrops, concealed by late Tertiary and Quaternary deposits in the Rio Grande Depression. To the northeast, the system of two slightly divergent faults is concealed by Quaternary gravels peripheral to the San Pedro Mountain. However, the system is believed to be continued east of the Ortiz Mountains by the Los Angeles and Las Jollas fault systems in the Galisteo structural basin.

Galisteo structural basin.—The area between the Ortiz Mountains and Glorieta Mesa is a broad structural basin with a gentle northward plunge. Igneous rocks have invaded the west limb of the basin in the Ortiz Mountains and Cerrillos Hills (Stearns, 1952b), interrupting the general easterly dips still evident west of the hills in the Galisteo Monocline. The synclinal trough is defined in a general way by outcrops of the Galisteo formation near Omero and of Espinaso volcanics where they impinge on Galisteo Creek. However, the general basin structure is interrupted by transverse faulting and folding, and a well-defined axial trace cannot be followed continuously through that distance. The east half of the basin is broken by faults related to the fault boundary of Glorieta Mesa.

Los Angeles fault system.—The Los Angeles fault system, trending north-easterly across the Galisteo structural basin, is marked by stratigraphic displacement, by sharp local deformation of beds adjacent to individual faults or fault zones, and commonly by strong dissimilarity of local structural pattern in beds on opposite sides of the fault zone. Individual members of the system appear to have a linear or gently sinuous trace, and are probably steep, but the dips of fault surfaces are not known from direct observation.

The fault system is most clearly demonstrable in the valley of the Arroyo Los Angeles northeast of Kennedy, and is therefore given that name. Here (Loc. L, Fig. 7), a narrow slice of complexly folded beds, ranging in age from uppermost Morrison (?) to lowermost Carlisle (all mapped as lower Mancos Shale in Pl. 1), separates gently-dipping beds of the upper Galisteo formation to the northwest from steeply-dipping beds of upper Mancos shale and Galisteo formation to the southeast. The latter beds are disposed in a sharp, assymmetric syncline, the axial trace of which parallels the south boundary of the fault slice. Northwest of the fault slice, uniform westerly dips prevail in a wide area, except in some exposures near the fault slice. These homoclinal beds are broken by steep (gravity?) faults trending N20°W to N30°W, at least two of which terminate against the fault slice. These relationships imply that the fault slice separates two blocks which were deformed internally, on independent lines, at the time of faulting. The oldest beds exposed northwest of the fault slice include beds of limestone conglomerate, which are believed to form a persistent horizon near the middle of the Galisteo formation (Stearns, 1943, p. 307, Fig. 5). Southeast of the fault slice, limestone conglomerate does not occur and the beds exposed must lie entirely below that horizon. The beds exposed on either side of the fault slice therefore belong to mutually exclusive sections of the Galisteo formation, and the relative displacement between them is down to the northwest.

The fault slice has the structural relations of a horst, actively raised in a gravity fault zone of relatively small net displacement. The sharp syncline south of the fault slice contrasts strongly with the gentle homoclinal dips north of the fault slice; the compression indicated by the syncline suggests that the south boundary of the fault slice may be a thrust fault. None of the relationships appear to be compatible with uniform tension or compression normal to the fault zone; they are more reasonably explained by the compression of an irregular slice in a zone of horizontal shear.

The same net displacement, downward to the northwest, must continue within the Galisteo formation southwestward from the fault slice to Galisteo Creek, for limestone conglomerates have not been found in this inter-

vening area and an indeterminate thickness of beds including them is omitted. Displacement in this section of the area may be considered either a continuation of the net shift in the zone containing the slice of Cretaceous beds in the Arroyo Los Angeles, or only a continuation (with decreasing displacement) of the fault forming the northern boundary of the slice.

The position of the fault is probably marked near Kennedy by a complex offset in the El Creston dike (Pl. 1). East of Kennedy, El Creston is a prominent ridge dominating the north end of the Galisteo Lowland. Its general trend is regular, but three prominent water-gaps in the ridge are associated with offsets in the general trend. In Galisteo Creek and the Arroyo de Los Angeles, nearly continuous exposures show that the ridge-forming dike is continuous and not broken by faulting. North of Kennedy, on an extension of the fault, the dike is offset about 300 yards and continues to the west as the northerly of a pair of parallel dikes about 200 yards apart. Exposures are not continuous in the area of offset, but both east and west of the offset the dike swings northward in the portions exposed.

Coincidence of the area of offset with the approximate position of a fault demanded by stratigraphic evidence implies that the offset takes place at a fault zone. The complex pattern of offset could be produced by a younger fault having a large strike-slip component, but the same pattern could develop, and the absence of one dike east of Kennedy would be less remarkable if the dikes were intruded after the Los Angeles fault system was formed.

Exposures are poor between the northeast end of the fault slice and Lamy; beds of the Galisteo formation crop out in a generally synclinal structure, but details cannot be demonstrated. East of Lamy, the northeast-trending canyon of Galisteo Creek follows a faulted monocline between the Sangre de Cristo Mountains and Glorieta Mesa. Two miles east of Lamy, the Glorieta sandstone and overlying rocks abut Precambrian rocks on a northeast-trending gravity fault (Fig. 7; Pl. 1). Darton (1928a) mapped the fault continuously for 10 miles to the northeast, but Read *et al* (1944) show the structure as principally monoclinal. In either case, the faulted monocline lies approximately on a projection of the Los Angeles fault system. Displacement is in an opposite sense, downward to the south, but geometric coincidence implies that the structure northeast of Lamy continues the same northeast-trending line of weakness.

Continuity of the Los Angeles fault system southwest of Galisteo Creek is postulated from the following observations:

(1) Vertically-dipping Cretaceous shales crop out in a small gulch 1½ miles southwest of Galisteo Creek (Loc. K, Fig. 7) as a linear slice, separating a syncline in the Galisteo formation to the southeast from an anticline in the Galisteo formation to the northwest.

(2) At the edge of the Ortiz pediment 3 miles southwest of Galisteo Creek (Loc. J, Fig. 7), a radial porphyry dike striking N80°W is offset through a distance of 300 yards in which the dike trends N45°E and dips 75°NW. A branch dike continues northeastward in the direction of offset. Beds of the Galisteo formation on both sides of the dike dip northeastward; they are different lithologically, but stratigraphic displacement cannot be demonstrated. A northeast fracture of some importance is suggested, antedating intrusion of the radial dike.

(3) Steeply-dipping sandstone and red clay of the Galisteo formation crop out locally in a gulch cut in Quaternary gravel 4 miles southwest of Galisteo Creek (Loc. H, Fig. 7). The outcrop lies only a mile west of exposures of the Mesaverde formation and the Pelon sill, striking N70°W and dipping 15°N. Complex structure is demanded in the intervening area; a fault thrown down to the northwest would be entirely appropriate.

None of the observations is conclusive, nor can displacement in the intervening areas be demonstrated. However, all three localities lie on a projection of the trace of the Los Angeles fault system; their alignment suggests continuity.

A parallel fault, 1½ miles to the northwest, is postulated from the following observations:

(1) Cretaceous shales are exposed in a small dome 3 miles northeast of Dolores (Loc. F, Fig. 7), along the edge of the Ortiz pediment. Its west boundary, not well exposed, appears to be a fault, trending northeast and downthrown to the northwest.

(2) Two miles northeast of this locality (Loc. G, Fig. 7) limestone conglomerate of the

Galisteo formation forms a low hog-back striking northeast, and dipping 50° to 60° NW. The hog-back is intruded by an en échelon swarm of porphyry dikes; their individual trends intersect that of the hog-back at low angles. A line of small springs, commonly associated with faults in other parts of the area, emerges on the southeast side of the hog-back. At the mouth of the Arroyo Choro, the limestone conglomerate outcrop swings eastward, with increased dips, as a monoclinal interruption of generally gentle northward dips. The outcrop ceases abruptly at the flood-plain of Galisteo Creek; as noted above, the horizon does not reappear for 6 miles north of Galisteo Creek. The effect of the local monocline is similar to that of a cross-fault between the two inferred northeast-trending faults, and may be interpreted as a bifurcation of the single fault north of Galisteo Creek.

(3) A sharp synclinal axis (Dolores syncline, Fig. 7) parallels the inferred fault trace between Dolores and Galisteo Creek. Although concealed by Quaternary gravel in most of this distance, outcrops near Dolores suggest continuity that far. The syncline is best explained as interruption of general easterly dips by drag associated with the inferred fault.

The northeast-trending faults of the Los Angeles fault system are certainly continuous northeast of Galisteo Creek, and probably continuous as a double line southwest to the Ortiz Mountains, a total distance of 12 miles. The northeast-trending Tijeras fault system is continuous for 26 miles southwest of the Ortiz Mountains. It is probably no coincidence that the two fault systems can be joined by projection across the saddle between the main mass of the Ortiz Mountains and Lone Mountain, to the southeast. Section BB′ (Fig. 5) indicates that a fault with displacement downward to the northwest can reasonably exist, and may be required, in that saddle. The gap in exposures is 8 miles, but this distance is reduced by the inferred displacement in the saddle between the Ortiz and San Pedro Mountains. The Los Angeles fault system is therefore believed to be continuous with that part of the Tijeras fault system north of the San Pedro Mountains.

Zone of diapir folding.—Between the Pelon sill and Galisteo Creek, the Galisteo formation is complexly folded. Upper Cretaceous shales are exposed in five small domes, characterized by steep to vertical dips in a central area of shale and moderate dips in peripheral beds of the Galisteo formation. In several localities, steeply-dipping shales crop out so close to moderately-dipping sandstones as to suggest strong angular unconformity, not consistent with the rather gentle regional deformation postulated for pre-Galisteo time. It is therefore believed that the folds are diapiric, and that the discordance arises from active mobility of the shales in the domal structures. Between these areas, folds in the Galisteo formation are highly irregular. Indivdual axial traces persist only short distances, and abrupt changes of trend are common. The domes in which Cretaceous beds are not exposed may be diapiric at depth.

This isolated zone of complex folding is, so far as known, unique in the Galisteo area. The trends of individual anticlines are not systematic, and that of the zones as a whole is east-west, transverse to other known structural features. The zone lies in the trough of the Galisteo structural basin, as does the Pelon sill farther south. It also lies in the same general area as does the inferred bifurcation of the Los Angeles fault system south of Galisteo Creek. The presence of intrusive igneous rock in one of the diapirs (Loc. F, Fig. 7) suggests that all of the diapirs may be related to igneous intrusion at depth. However, a single, convincing explanation of the complex structure cannot be drawn from the data at hand.

Las Jollas fault system.—Beds of the lower Mancos shale crop out in a gentle dome[2] in the southwest portion of the Canyada Las Jollas, separated by a northeast-trending fault from nearly flat-lying beds of the upper Mancos shale exposed in the escarpment of the Estancia Valley. The fault separates beds of diverse structure as well as diverse age. Stratigraphic displacement decreases to the northeast, but northeast-trending faults of small displacement near Omero are believed to be part of the same fracture system. Similarly, northeast-trending faults in the Galisteo Lowland, asso-

[2] The dome appears to center on Lone Mountain and is presumed to be related to igneous intrusions in that vicinity, but the complete structure has not been mapped.

ciated with the fault boundary of Glorieta Mesa, to be described below, appear to continue the same fault system. These several faults are here called the Las Jollas fault system; their position and trend is appropriate to a continuation of that part of the Tijeras fault system south of the San Pedro horst.

Fault boundary of Glorieta Mesa.—The Galisteo Lowland is underlain by upper Cretaceous rocks dipping generally westward, in the east half of the Galisteo structural basin. Glorieta Mesa is underlain by Triassic and older rocks, gently warped but relatively flat-lying. The two areas are separated by a complex fault system, with net displacement downward to the west. An highly irregular topographic boundary, controlled by the relative resistance of rocks involved in the fault system, only imperfectly reflects the structural boundary.

Details of the fault system are not completely known. The structure of the pre-Cretaceous rocks was not mapped in detail. The faults mapped within the upper Cretaceous rocks are demanded by offset of the horizons traced in the field, but observed outcrops by no means provide complete continuity of those horizons.[3] In the area south of the Arroyo San Cristobal, where relatively abundant horizon-markers occur in the lower part of the upper Cretaceous series, the faults shown on Plate 1 and Figure 7 are well located and no important faults are omitted. In the valley of the Arroyo San Cristobal and northward as far as Lamy, available outcrops are not sufficient to yield more than a generalization of an unquestionably complex system. North of El Creston, outcrops were not found in the area mapped as, and presumably underlain by, upper Cretaceous rocks.

At Lamy, the saddle between Glorieta Mesa and Cerro Colorado, in which the Galisteo formation is exposed, marks the position of a fault trending N15°E, which appears to branch from the Los Angeles fault system. South from Cerro Colorado to El Creston, exposures of Cretaceous rocks west of the boundary fault are not known, and the western limit of Triassic outcrops is more or less irregular, suggesting that the fault pattern is complex.

Between El Creston and the Arroyo San Cristobal, a north-south fault separates upper Cretaceous rocks to the west from Triassic rocks of complex structure immediately to the east. A broad area without outcrops separates these Triassic beds from those in Glorieta Mesa proper. West of the boundary fault, the upper Cretaceous beds are broken by two antithetic faults, tending north-south.

South of the Arroyo San Cristobal, the Morrison (?) formation outcrops in a broad mesa, separated from Glorieta Mesa proper by the valley of the Arroyo Jaspé, and broken by numerous faults of small displacement. Only a few of these faults are shown in Plate 1; they are omitted from Plate 4. The area of Morrison (?) outcrop is separated from the Triassic rocks of the Glorieta Mesa proper by faults generalized in Plates 1 and 4. The fault inferred in the valley of the Arroyo Jaspé is exposed north of the Arroyo San Cristobal, in an unnamed tributary (Pl. 3, fig. 2). It is essentially vertical in this outcrop. However, the Morrison (?) beds partake in general of the relatively flat-lying structure of the Triassic beds farther east (Read *et al.*, 1944). Several en échelon faults, slightly concave to the northwest,[4] separate the mesa from the lowland eroded in upper Cretaceous rocks, and continue some distance into the lowland. Steepening of the otherwise gentle dips generally occurs adjacent to the faults. For about a mile of its length, the Arroyo Jara follows a monoclinal structure which provides a part of the same displacement down to the west.

The upper Cretaceous rocks in the Galisteo Lowland form the west half of a broad dome, the crest of which lies between the Arroyos Jara and San Cristobal. Its position is outlined by the top of the lower Mancos shale, marked by the outcrop and inferred position of the Juana Lopez member of the Mancos shale (Pl. 1). The east half of the dome does not appear to be present in the pre-Cretaceous rocks farther east. According to Read and Andrews

[3] The extent of control by actual outcrop is indicated generally in Plate 1 by the relative extent of solid and broken-line contacts. In fact, somewhat more complete control than shown was provided by supplementary horizon-markers, not included in Plate 1.

[4] The concavity is conspicuous in aerial photographs, and undoubtedly real.

(1944), low westerly dips prevail 3–5 miles east of the boundary of Glorieta Mesa. The dome is therefore a local interruption of a more general homoclinal structure, and is probably genetically related to the faulting. The dome is further broken by a prominent north-south antithetic fault, and by northeast-trending faults of lesser displacement, probably belonging to the Las Jollas fault system. The latter faults appear to terminate against the north-south fault, but this implied relationship may arise from ignorance; outcrops are lacking east of the north-south fault in the critical area. The north-south fault appears to be systematically offset eastward where it intersects the en échelon northeast fractures. However, the apparently systematic offset is not real, as displacement on the two northeast-trending fractures is in the opposite direction. The faults of diverse trend are not of diverse age, but belong to a single complex system formed contemporaneously.

South of the Arroyo Jara, displacement downward to the west is continued by faults trending nearly north-south as far as the limits of the area mapped. The Jurassic beds adjacent to the boundary fault dip moderately northwest and are brecciated by numerous small faults (Pl. 3, fig. 3; only two larger faults are shown in Pl. 1) striking east-northeast.

Thus, the fault boundary of Glorieta Mesa includes two sections trending north-south, offset between the Arroyos San Cristobal and Jara. The offset is coincident with the inferred termination of the Las Jollas fault system against the fault boundary of Glorieta Mesa; the trends of the off-setting boundary faults are intermediate between those of the two fault systems elsewhere. The two fault systems appear to be contemporary in the area of offset. Thus, although faults of the Las Jollas fault systems are of only small stratigraphic displacement at the intersection, the resultant pattern suggests an intersection of two fault systems of equal regional importance.

The upper Cretaceous rocks of the Galisteo Lowland are separated from nearly flat-lying beds in the scarp of the Estancia Valley by a northwest-trending fault of varying displacement downward to the southwest. The fault is transverse to both the Las Jollas fault system and the fault boundary of Glorieta Mesa. It is perhaps an adjustment in the acute angle bounded by those systems.

The age of the fault boundary of Glorieta Mesa cannot be established directly except that, near Lamy, it is post-Galisteo. None of the faults in the Galisteo Lowland could be traced northward into the Galisteo formation, and they presumably die out northward. One of the north-south faults crosses a wide gap in El Creston. Projections of the dike on either side of the gap indicate offset, but exposures of the dike are lacking for 200 yards. The relationship of the apparent offset to the fault is thus not established. The dike crosses the next fault east without visible offset, and terminates abruptly at or near the boundary fault between Cretaceous and Triassic rocks. The north-south faults, and presumably the entire system, are therefore believed to antedate intrusion of the dike.

Summary and interpretation.—Deformation in the interval post-Espinaso and pre-Abiquiu (?) outlined the following structural features:

(1) A northeast-trending fracture system, including the Tijeras fault system south of the Galisteo-Tonque area, the Los Angeles and Las Jollas fault systems in the Galisteo structural basin, and the fault system separating the Sangre de Cristo Mountains from Glorieta Mesa. Although absolute continuity cannot be established, gaps are relatively small and regional failure on a line of weakness more than 60 miles long is indicated. The fracture system separates blocks which differ in both general and local structural pattern. Fault surfaces are steep and generally normal. However, relative displacement varies on strike and one important fault has been described as a thrust. Furthermore, deformation within the fault slice exposed in the Arroyo Los Angeles and in beds immediately southeast of it suggest that the fault slice moved upward by compression in a "normal" fault zone. Neither uniform tension nor uniform compression normal to the fracture system is suggested, and the fracture system is tentatively interpreted as a horizontally directed shear. Strike-slip cannot be demonstrated, but rotational stress in the horizontal plane is implied.

(2) The block southeast of the Tijeras-Las Jollas fault systems, largely concealed by Quaternary gravels in the Estancia Valley, is

broadly synclinal, characterized by gentle dips in the Manzanita Mountains to the west and the southern extension of Glorieta Mesa to the east. The pattern is interrupted locally by faults of small displacement. The fault boundary of Glorieta Mesa and probably the monocline east of the San Pedro horst extend for some distance beyond the Las Jollas fault system into the southeast block, and the igneous rocks of South Mountain are a southward extension of the line of intrusions north of the Tijeras fault system. In general, however, the structure is gentler and simpler than that northwest of the Tijeras fault system. Reiche (1949, p. 1210) has recently commented on the absence of "Laramide" (pre-Abiquiu (?)) deformation in the Manzanita Mountains.

(3) The area northwest of the Tijeras and Los Angeles fault systems is thrown into broad folds, plunging northward. Anticlinal axes lie in the Sandia Mountains, the Tuerto faulted anticline, and possibly in the Sangre de Cristo Mountains east of the area mapped; synclinal axes in the Tonque Valley and Galisteo structural basin. As indicated in section BB′ (Fig. 5), the Tuerto faulted anticline may be regarded as a faulted anticline, a faulted monocline, or a tilted block bounded to the west by flexure as well as faulting. As inferred by Darton (1928a, p. 99–100), the Sandia uplift was probably a faulted anticline, but the west limb has been cut off by post-Santa Fe faulting. Faults in the steep west limbs are normal, and do not imply compression. That the gentler east limbs of both folds were subject to tension, is indicated by the grabens east of the Sandia axis (Grabens A, B, and C, Fig. 7), and implied by the north-south line of intrusions in the Ortiz Mountains and Cerrillos Hills. The tensional faults may be ascribed to relaxation in folds formed by earlier compression; or the composite structures may be interpreted as the adaptation of shallow-seated beds to folding by compression of a deep-seated layer. The writer prefers the latter interpretation, first advanced by Gilbert (1875, p. 62) and supported by Gilluly (1932, p. 88) for Basin Range Structure. Deep-seated, east-west compression is compatible with northeast shear in the Tijeras and Los Angeles fault systems.

The La Madera faulted anticline and minor folds in the Tonque Valley trend northeast, transverse to the north-south structures and parallel to the Tijeras fault system. Their position is between the Sandia and Tuerto faulted anticlines and opposite that portion of the Tijeras fault system which has been described as a thrust fault. It is tentatively suggested that the northeast-trending folds reflect northwest compression in the footwall of the steep thrust fault.

The Galisteo structural basin shares the north-south trend of the Sandia and Tuerto faulted anticlines, but is broader and gentler, implying that intensity of deformation died out northeastward. Horst and graben structure, outlined by northwest-trending faults, characterizes the east limb of the structural basin. Like the grabens of the Tonque Valley, the faults may indicate stretching in the west limb of an anticline raised in the Sangre de Cristo Mountains, east of the area mapped.

(4) The slightly divergent faults of the northeast fracture system enclose a block of varying displacement relative to adjacent areas, the several sections separated by transverse faulted monoclines. Thus, the Tijeras graben is separated from the San Pedro horst by a sharp, transverse monocline, and the San Pedro horst must in turn be separated from the gently synclinal Galisteo structural basin by a transverse monoclinal structure concealed by Quaternary gravels. The faulted monocline forming the west boundary of Glorieta Mesa is a similar transverse structure, against which the southeastern member of the northeast fracture system terminates. In a general way, the Tijeras graben, San Pedro horst, and Galisteo structural basin within the fault block correspond in position to the Uña de Gato basin, Tuerto faulted anticline, and Galisteo structural basin north of the Tijeras and Los Angeles fault systems. The north-south line of intrusions cuts completely across and beyond the fault block. Thus, the wedge-shaped fault block shares its principal structural elements with those to the north. Complete continuity does not obtain, however; the Tijeras and Los Angeles fault systems are accompanied by numerous transverse structures, such as the La Madera faulted anticline, developed by independent internal deformation of the adjacent blocks and interrupting the major north-south trends.

Late Tertiary History of North-Central New Mexico

Introduction

Deformation and erosion in the interval post-Espinaso and pre-Abiquiu (?) established a pattern upon which late Tertiary deformation and deposition were superimposed. The Galisteo-Tonque area is marginal to the Rio Grande Depression, and its late Tertiary history is only part of that of the larger structural entity. Because of recent emphasis on the complexity of the late Tertiary Albuquerque-Belen basin (Wright, 1946), discussion of the late Tertiary evolution of the Galisteo-Tonque area is extended to include available data on the Rio Grande Depression in northern New Mexico.

Abiquiu Basin of Deposition

The lithologic character of the Abiquiu (?) formation in the Tonque Valley and in the La Bajada escarpment may be taken to indicate the following general conditions of deposition:

(a) The nontuffaceous beds of the formation could have been derived entirely from erosion of the Galisteo and Espinaso formations. Mesozoic rocks, which break down principally into quartz sand and clay, or Tertiary intrusives, generally indistinguishable in hand specimen from similar pebbles in the Espinaso volcanics, could have contributed débris without appreciably altering the character of beds. However, Precambrian rocks; the Magdalena group, from which at least chert would have survived; Permo-Triassic red beds, of which some, at least, of the color would persist in the sand sizes; and contact metamorphic rocks, common in the Cretaceous shales and older rocks of the Ortiz-South Mountain chain, were not recognized. Their absence implies that erosion of the highlands produced by pre-Abiquiu (?) deformation had not stripped great thicknesses of their original cover.

(b) The fine-grained detritus implies that near-by areas had no considerable relief. The Espinaso volcanics, at least, are a potential source of coarse clastic fragments. Increase of grain size in the upper part of the Arroyo Pinovetito section and comparable coarsening of materials overlying the tuffs north of La Bajada indicate some accentuation of relief in the latter part of the sedimentary cycle.

(c) Decrease in grain size northward in the Tonque Valley, and between the Tonque Valley and the La Bajada escarpment, suggests (1) a northerly drainage out of the area of the present Tonque Valley, from source areas to the south, and that, (2) if source areas existed to the east or west, their relief diminished rapidly from south to north. The latter inference is in accord with the northward plunge of pre-Abiquiu (?) structures west of the Ortiz Mountains.

(d) The tuffs of the La Bajada escarpment were presumably carried into the area by southward-flowing streams heading far to the north in the Taos plateau.

(e) The thin-bedded limestones, and probably the tuffs, of the La Bajada escarpment were deposited in bodies of standing water. A lake did not occupy the area continuously, but varied in extent or position seasonally, or as determined by the variable accumulation of sediment. However, a lake or lakes must have existed in the vicinity during much of Abiquiu (?) time.

(f) The consistent color difference between the Abiquiu (?) and Santa Fe formations may reflect a difference in the climatic conditions governing weathering, in the nature of the rocks exposed, or both. The common decomposition of porphyry pebbles in the Abiquiu (?) formation, in contrast to the relative freshness of similar pebbles in the Santa Fe formation suggests that chemical weathering was more effective in Abiquiu (?) time.

In the Tonque Valley the base of the Abiquiu (?) formation generally overlies the Espinaso volcanics. In the southern part of the Tonque Valley, however, the formation bevels a minor fold in the early Uña de Gato Basin and laps across part of the Tuerto faulted anticline. These structures were formed in pre-Abiquiu (?) time, although renewed movements in the fault zone may have occurred in post-Abiquiu (?) time. The total shift in the Tuerto faulted anticline, downthrown to the west, is about 7000 feet (Section BB′, fig. 5). The initial fault scarp was destroyed before or during Abiquiu (?) time. It is probable, however, that the scarp retreated eastward, and that a low highland occupied the site of the present Ortiz

Mountains during Abiquiu (?) time. Erosion of the Espinaso volcanics and Galisteo formation in this highland would undoubtedly have exposed the upper portions of the larger intrusive bodies and, by greater resistance to erosion, the intrusive rocks would have formed hills then as now. As erosion stripped the sedimentary cover, the size and perhaps even the relief of the hills would increase. The increase of grain size higher in the Abiquiu (?) formation may reflect such increasing exposure. It is not sufficient to demand a large increase of relief.

A highland north of the present Sandia Mountains is indicated by stratigraphic relations at Placitas. Assignment of fine-grained beds in the Norins Well to the Abiquiu (?) formation implies that the near-by highland was low and may indicate that the Ortiz-South Mountain highland east of the Tonque Valley was a more effective source of detritus than the Sandia highland to the west.

The Tonque Valley, therefore, was the site of a local basin of deposition, bounded on the east by an eroded scarp and on the west by slopes cut into a tilted block modified by faulting. It was probably further bounded on the south by scarps associated with the La Madera faulted anticline. None of the bounding highlands had great relief; that to the east may have had greater relief than that to the west. Streams flowed northward through the basin of deposition toward lakes in the vicinity of La Bajada.

The basin of deposition may well have been continuous between La Bajada and the Arroyo Hondo. Present exposures of Espinaso volcanics east of the trace of the Tuerto faulted anticline (4 miles south of Rosario Siding, Pl. 1; Loc. E, Fig. 7) imply that the magnitude of that fold decreases northward and that it probably dies out altogether. The margin of the sedimentary basin would have swung eastward some distance south of Rosario Siding, probably to a northwestward-facing scarp associated with the Los Angeles fault system east of the Ortiz Mountains. In the Arroyo Hondo section, both the absence of a thick tuff horizon derived from the north and the probable derivation of materials from erosion of the Galisteo formation are consistent with northwestward drainage from such a scarp. The apparent thinning of the Abiquiu (?) formation between the Arroyo Hondo and Santa Fe probably indicates the approach to greatly reduced remnants of highlands raised in the Sangre de Cristo Range in pre-Abiquiu (?) time. The thinning might be interpreted as the result of post-Abiquiu (?), pre-Santa Fe erosion consequent upon uplift of the Sangre de Cristo Range. However, Denny (1940b, p. 690–691) has shown that the Santa Fe basin of deposition included much of at least the western half of the Sangre de Cristo Mountains, which implies that pre-Santa Fe uplift did not take place near Santa Fe. Certainly, the relief of the Sangre de Cristo Range was not great in Abiquiu (?) time. It may, however, have remained an area of erosion rather than of sedimentation.

The areas described were, of course, re-entrants of a much larger sedimentary basin in which the Abiquiu (?) formation was deposited. The pattern of the re-entrants corresponds generally to that of structures described in an earlier section as pre-Abiquiu (?). Overlap of the Tuerto faulted anticline by the Abiquiu (?) formation implies that most of these structures are pre-Abiquiu (?). The correspondence of pattern obtains because eroded remnants of older structures existed on the margins of a new, subsiding basin of deposition. Deformation contemporary to the Abiquiu (?) formation is implied by the recorded change from erosion to deposition, by maintenance of a supply of detritus to areas of deposition, and by variation in thickness of the Abiquiu (?) formation. However, this intra-Abiquiu (?) deformation followed new lines, indicated by a paleogeographic sketch of the sedimentary basin in north-central New Mexico (Fig. 8).

The Abiquiu tuff fan radiated from contemporary volcanic centers probably located in the present Taos Plateau (Butler, 1946, p. 81–87). Detritus from these centers flooded areas of considerable relief in the Tusas quadrangle and spread southward into and beyond the Abiquiu quadrangle. In the southern half of the Abiquiu quadrangle, the volcanic débris spread out on an erosion surface of low relief (Smith, 1935, p. 949), which extended southwest at least as far as San Pedro Mountain, where it bevelled the early Tertiary Nacimiento uplift (Church and Hack, 1939). The

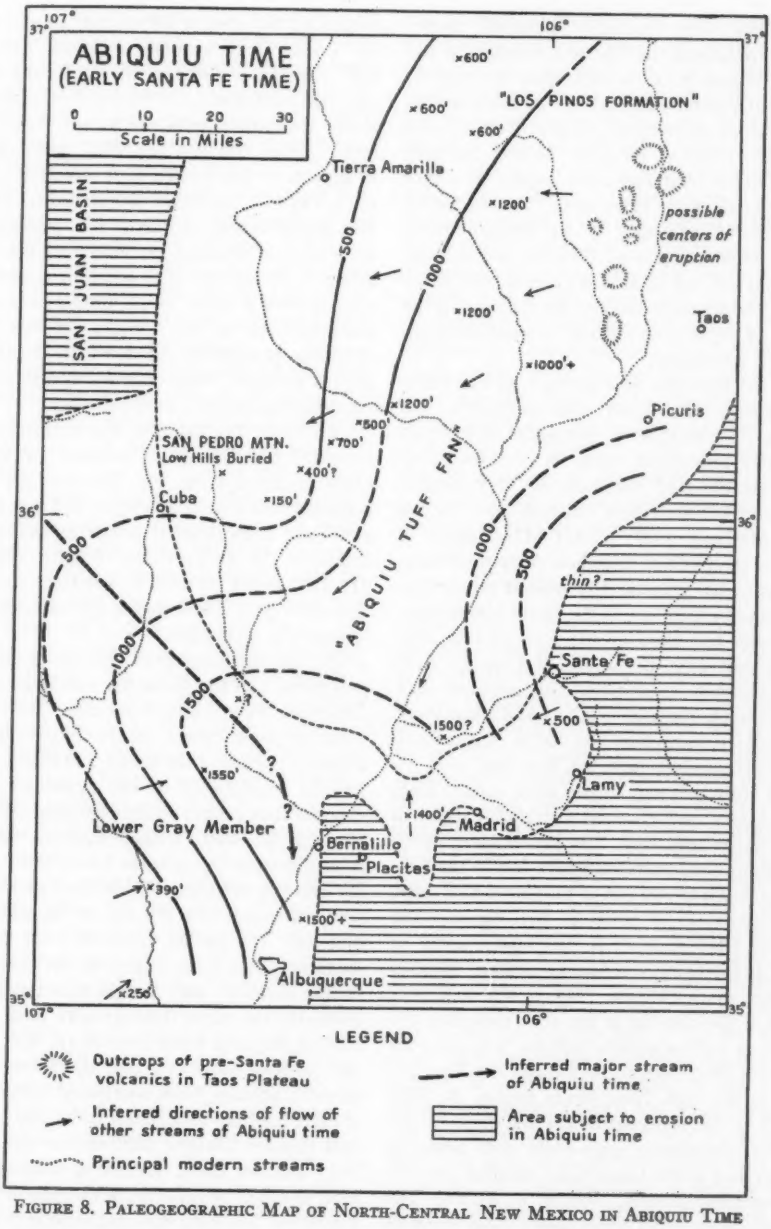

FIGURE 8. PALEOGEOGRAPHIC MAP OF NORTH-CENTRAL NEW MEXICO IN ABIQUIU TIME

few monadnocks rising above the surface in San Pedro Mountain were probably buried by the Abiquiu tuff.

Subsequent erosion has removed the erosion surface and overlying deposits from the southern end of the Nacimiento uplift, but similar

conditions very probably obtained there. Smith considered the Chicoma volcanics of the Valles Mountains to be older then the Abiquiu tuff, but pointed out that the younger formation carries no evidence of contribution from the older (1938, p. 950–952). There is, in fact, no evidence that highlands existed in either the Nacimiento or Valles areas. The southward decrease of grain size in the Abiquiu quadrangle implies that none did exist, and that the Abiquiu tuff fan blanketed at least part of the area. The southern limit of the fan, and of the basin of deposition, did not lie north of those present mountains.

In the Rincones de Zia, south of the Sierra Nacimiento, the Lower Gray member of the Santa Fe formation is distinguished by consistently gray-white color. It is overlain by the Middle Red member, of very similar lithology but strikingly different color (Bryan and McCann, 1937, p. 811–812). Lenses of pebbly sand in both members carry small fragments of red granite. The nearest present exposures of granite are in the Sierra Nacimiento, probably too close at hand to provide only small fragments. Bryan and McCann (p. 814–815) therefore suggest that the fragments may have been introduced by a long and roundabout drainage system from the Sierra Nacimiento. It is also possible, however, that they were derived from outcrops in the more distant Zuni Mountains, and that no highlands existed in the Nacimiento area. The Lower Gray member thins southward from 1551 feet in the Rincones de Zia to 390 feet in the Canyada Pilares, a distance of 10 miles (p. 816, fig. 5). The Lower Gray member is somewhat coarser as well as thinner in the Apache Graben, 10 miles further south (Wright, 1946, p. 402–403). The bulk of the material in this area, including volcanic as well as granitic pebbles, was probably derived from Mount Taylor and the Zuni uplift to the west. Near Jemez pueblo, 10 miles north of the Rincones de Zia and closer to the Sierra Nacimiento, gray-white beds probably equivalent to the Lower Gray member are very fine-grained (Renick, 1931, p. 56, Pl. 7A). The combination of all these observations suggests that the Lower Gray member was deposited by streams flowing northeastward into a subsiding basin northeast of the Rincones de Zia.

It is consistent with, although by no means demanded by, present information to correlate the Lower Gray member with the Abiquiu tuff. The Lower Gray member, like the Abiquiu tuff, underlies more reddish beds carrying fossils compatible with the known Santa Fe fauna (Bryan and McCann, 1937, p. 807–808). It appears to date from a time when highlands did not exist in the Sierra Nacimiento or Valles Mountains—when, in fact, the position of these present highlands lay within a basin subsiding more rapidly than adjacent areas. The axis of such a basin would provide a logical southern limit to the Abiquiu tuff fan. The axis was presumably occupied by a master stream to which both northward- and southward-flowing streams were tributary.

This basin lies athwart the course of the ancestral Rio Grande as outlined by Bryan (1938, p. 205–208, fig. 46). However, the river certainly post-dates the Abiquiu tuff fan, which continued uninterrupted across its postulated course in the Abiquiu area. Wright (1946, p. 411–412) shows that the Lower Gray member may date from a time before the ancestral Rio Grande was established.

The age relationship of the sub-basin here postulated and the Gabaldon sub-basin of the Lower Rio Puerco area is not clear. The Lower Gray member cannot be recognized in the southern portion of the Lower Rio Puerco area, and its relationship to the Gabaldon playa deposits thus cannot be demonstrated (Wright, 1946, p. 402–403). Wright suggests that the color difference between the Lower Gray member and the overlying Middle Red may disappear southward with changes in the sources of material. The darker (reddish) beds farther south may thus be equivalent to both members farther north (p. 402). Wright thus tentatively correlates the playa deposits with both members of the area farther north (p. 410). It is also possible, however, that the Lower Gray member actually thins southward rather than grading laterally into darker beds. The Middle Red member thickens southward as the Lower Gray member thins, but not at a comparable rate, as might be expected in the case of lateral gradation. The color differentiation may be the product of climatic differences rather than of differing source materials. It is here postulated, therefore, that the Lower Gray member is thin or absent in the Lower Rio Puerco area, and

that the Gabaldon playa deposits are entirely equivalent to the Middle Red member. The fossil content of the Middle Red member indicates that it is equivalent to part of the Santa Fe formation in the type locality rather than to the Abiquiu tuff. The principal development of the Gabaldon playa may therefore post-date the beds here correlated with the Abiquiu tuff.

Available data on present thicknesses of the beds here correlated with the Abiquiu tuff are plotted in Figure 8, and generalized isopachs have been sketched. Present thicknesses, of course, are the combined results of irregularities in the pre-Abiquiu topography, of warping contemporary to deposition, and of subsequent erosion. However, most of the thicknesses cited were obtained in sections overlain by the Santa Fe formation, and should approximate original thicknesses.

The available data suffice to show that the basin of deposition was complex. A rather symmetrical trough north of Santa Fe is suggested by (1) the isopachs as drawn, although data on the east side are nearly lacking and may constitute evidence of post-Abiquiu (?), pre-Santa Fe erosion; and (2) the prominent southward extension of the Abiquiu tuff fan. A second basin south of the present Valles Mountains, oblique to the first, is implied by the considerable thicknesses in the Rincones de Zia, Tonque Valley, and New La Bajada Hill. The southeast margin of the basin was irregular because of pre-existing irregularities of topography.

It may be presumed that the axis of this subsidiary basin was the site of an important stream, carrying convergent drainages from the northeast, northwest, and southwest toward the deep portion of the basin. The fate of such a stream is conjectural. It may have ended within the basin; the occurrence of lacustrine deposits at New La Bajada Hill suggests that the basin was closed. However, the lakes may have been formed locally at times when active aggradation of the Abiquiu tuff fan impinged upon the north end of the Sandia Highland, blocking and temporarily ponding drainage from the Tonque Valley and Galisteo Basin. Hence, it is also possible that the Abiquiu basin, like the later Santa Fe basin, drained southward west of the Sandia Mountains.

The Norins Well lies in a position critical to the problem, and convincing identification of the Abiquiu (?) formation in its log would perhaps provide a solution. Certainly, if the Abiquiu (?) formation is represented in the upper, coarse-grained beds, its thickness (a fraction, probably small, of 2150 feet) is less than that in the Rincones de Zia and in the La Bajada escarpment. Closure of the isopachs on the southern margin of the postulated sub-basin and probable confinement of the drainage would be strongly suggested. However, interpretation of some part of the fine-grained gray beds as representing the Abiquiu (?) formation would suggest a thickness comparable to those farther north, perhaps even greater. Active subsidence of an area west of the Sandia Mountains would be required (but without effective accentuation of the Sandia highland) and maintenance of through drainage to the south implied. The latter interpretation is considered most consistent with the data at hand.

Santa Fe Basin of Deposition

The river deposits of the Galisteo-Tonque area have been interpreted previously (Bryan, 1938, p. 207; Denny, 1940b, p. 690; Bryan and Upson, unpublished manuscript) as part of the series of outcrops which more or less defines the course of the ancestral Rio Grande, which integrated several sedimentary basins in Santa Fe time. This river, like its modern counterpart, flowed through the west side of and west of the Galisteo-Tonque area. This area, therefore, was in the east side of the sedimentary basin in which the Santa Fe formation was deposited. The course of the ancestral Rio Grande swung through a broader belt than does the present Rio Grande. Within the Galisteo-Tonque area, river deposits and the upper part of the fan deposits west of the San Francisco fault are contemporaneous. Specific correlation of horizons is not possible, and faults break the areal continuity, but intercalation of the two types occurs. The Santa Fe lavas of the Mesa Santa Ana are also contemporary to river deposits. Accumulation of lava in that area would tend to push the ancestral Rio Grande eastward. However, the river maintained its course close to the lava margins, because of aggressive fan construction by its tributaries to the east.

The fan deposits west of the San Francisco fault grow coarser to the east (Fig. 4), further from the ancestral Rio Grande and closer to their source. The abundance of porphyry pebbles in gravel lenses implies an unrestricted drainage from the east across the San Francisco fault. Some pebbles of contact metamorphic rock types can only have been derived from the Ortiz-South Mountain chain. A heterogeneous mixture of igneous, sedimentary, and metamorphic rocks is exactly what erosion of these mountains does and would have produced. The drainage barrier which now exists east of the San Francisco fault between Placitas and the Arroyo Tonque cannot have existed in Santa Fe time. Streams must have flowed at will from an Ortiz highland west and northwest to the ancestral Rio Grande.

The lack of a drainage barrier east of the San Francisco fault does not imply that the fault was not active during Santa Fe time. North of the Arroyo Tonque, where the fault now brings Cretaceous and early Tertiary rocks to the surface, no drainage barrier exists at present. Similar conditions could have existed farther south in Santa Fe time. The calcareous sediments of the Mesa de la Cal are best interpreted as spring deposits associated with the San Francisco fault. Modern springs are associated with this and other faults in the area. Similar deposits, of Pleistocene and Recent age, blanket the south slope of the mesa at Alamos Altos ranch. Calcareous silts and thin lenses of travertine are interbedded with the youngest alluvium in the Arroyo Alamos Altos where it crosses the fault. It is believed that the travertines of the Mesa de la Cal indicate that the San Francisco fault was active during at least the last stages of deposition of the Santa Fe formation. It is probable, however, that a continuous mantle of sediment extended from the Ortiz Mountains to the ancestral Rio Grande.

The basal gravels at Placitas, considerably coarser than any observed elsewhere in the Galisteo-Tonque area, were presumably deposited close to their source. Gravel lenses containing porphyry pebbles are intercalated in these coarse beds, but are not the coarsest materials present. The large blocks are all red sandstone, except for a few blocks of Cretaceous sandstone at the very base. The porphyry pebbles, like those in other parts of the Santa Fe formation, originated in the Ortiz-South Mountain chain. The red sandstone blocks came from a highland on the site of the present Sandia Mountains, but capped by Permo-Triassic red beds, in contrast to the present Magdalena limestone cap. Much of the red bed sequence was probably eroded from the Sandia Mountains in Santa Fe time. It is also probable that the highland block, then as now, was bounded by the northeast-southwest faults near Placitas. Relief was probably not as great, however. The modern streams carry some blocks of Magdalena limestone and Precambrian granite at least twice as great as those of red sandstone in the Santa Fe formation.

The coarse blocks of red sandstone at Placitas occur in gravels interbedded with gravels containing rocks derived from the east. The highland to the south was therefore contributing to an already established drainage from the east. North of Placitas, and higher in the section, gravel lenses containing porphyries alternate with gravel lenses containing only sedimentary and crystalline rocks. It is believed, therefore, that the Sandia highland was uplifted at the beginning of this time sequence without disrupting an easterly drainage already established. As the bulk of the material contributed by the Sandia highland became greater (and the relief of the highland diminished), a northerly drainage from that mass alternated with the original drainage from the east in building fans. The failure to divert the original easterly drainage implies that relief at the north end of the Sandia highlands was never great. As the coarse gravels are found only at the base of the formation, the Sandia highland probably possessed maximum relief at the beginning of Santa Fe time.

The Santa Fe formation east of the Espinaso Ridge is probably at least in part equivalent to the fan deposits west of the San Francisco fault. In the former locality, the formation is somewhat coarser than in the latter locality. A heterogeneous pebble content, in which porphyries predominate, indicates derivation from erosion of a highland on the site of the Ortiz-South Mountain chain. The increase in content of coarse material and the lesser sorting than that of the Abiquiu (?) formation imply that relief of those hills was accentuated at the

beginning of Santa Fe time. There is no indication of further accentuation at a later date.

To the north, the proportion of porphyry pebbles in gravel lenses decreases. The fans of the Galisteo-Tonque area merge with those of the Santo Domingo Valley, which were fed by erosion of the Sangre de Cristo Mountains (Bryan and Upson, unpublished manuscript). The Ortiz Mountains were probably a peninsula-like highland extending northward into the sedimentary basin of Santa Fe time. It has not been established that hills occupied the site of the present Cerrillos Hills, but it is quite possible.

Thus, the Santa Fe formation in the Galisteo-Tonque area was deposited in a basin of somewhat different geography than that postulated for Abiquiu time. However, deposition still occurred in relatively small re-entrants of a much larger basin. The Ortiz and Sandia highlands were uplifted and acquired appreciable relief. It is believed that this rejuvenation was accomplished in large part by normal faulting on lines coincident with the known post-Santa Fe faults. Part of the Sandia block at and north of Placitas, a "highland" in Abiquiu time, was incorporated in the sedimentary basin. The formerly northward drainages of the Tonque Valley and Santa Fe Plateau were diverted west and southwest. The sedimentary basin of which the Galisteo-Tonque area formed part was eventually integrated with adjacent basins by the ancestral Rio Grande.

The Santa Fe formation of the Galisteo-Tonque area is to some extent equivalent to those parts of the formation in other areas which were deposited by steams graded to the same river. This includes the Upper Buff member of the Albuquerque-Belen Basin (Bryan and McCann, 1937, p. 815–817; Wright, 1946, p. 404–407) and equivalent beds in the San Acacia area which have yielded upper Pliocene fossils (Denny, 1940a, p. 93; Needham, 1936). The existence of the ancestral Rio Grande at the time of deposition of beds containing the type Santa Fe fauna has not been rigorously established. Wright's discussion (1946, p. 411–412) of the Gabaldon playa deposits, which are at least in part equivalent to beds carrying a compatible fauna, raises doubt of its existence at that time.

Reconstruction of the geography of the Santa Fe basin of deposition in north-central New Mexico (Fig. 9) is a composite of at least two discrete intervals, distinguishable only in the western border of the Albuquerque Basin. Here, data on thicknesses referable to two members (Middle Red and Upper Buff) imply some independent variation. However, data from other areas, all derived from partial or estimated sections, cannot be so apportioned, nor is their correlation with either or both of the bipartite divisions established. Hence, the isopachs sketched have no strong justification. Because the Upper Buff member in the Albuquerque Basin is generally thin, the suggested thickening from west to east would probably be supported by more adequate data. Certainly, thick sections are now confined to the eastern half of the basin. The thickening is the combined effect of intra-Santa Fe warping, intra-Santa Fe erosion, and post-Santa Fe erosion. The relative importance of each cannot be evaluated.

The Middle Red member probably represents the larger share of Santa Fe time. In the Albuquerque-Belen Basin, it is the thicker of the two members (conspicuously so south of the area included in Fig. 9), and scant fossil evidence suggests its equivalence to beds containing the type Santa Fe fauna (late Miocene to middle Pliocene), while the Upper Buff member and associated river deposits may represent only late Pliocene time. However, the geographic pattern shown in Fig. 8 is dominated by features which can be demonstrated only for late Santa Fe time. These include:

(a) Integration of the sedimentary basin by a through-flowing stream, the ancestral Rio Grande. In the southern Albuquerque-Belen Basin, by direct tracing into the Upper Buff member, and still farther south, by fossil evidence, the deposits have been dated as late Santa Fe.

(b) Probable uplift of a highland area in the Sierra Nacimiento, indicated by the lithologic change between Middle Red and Upper Buff members in the Rincones de Zia.

(c) Contemporary eruptions of rhyolitic lavas in at least the southern Valles Mountains (E. S. Larsen, Jr., private communication). The Peralta tuff member in the Santo

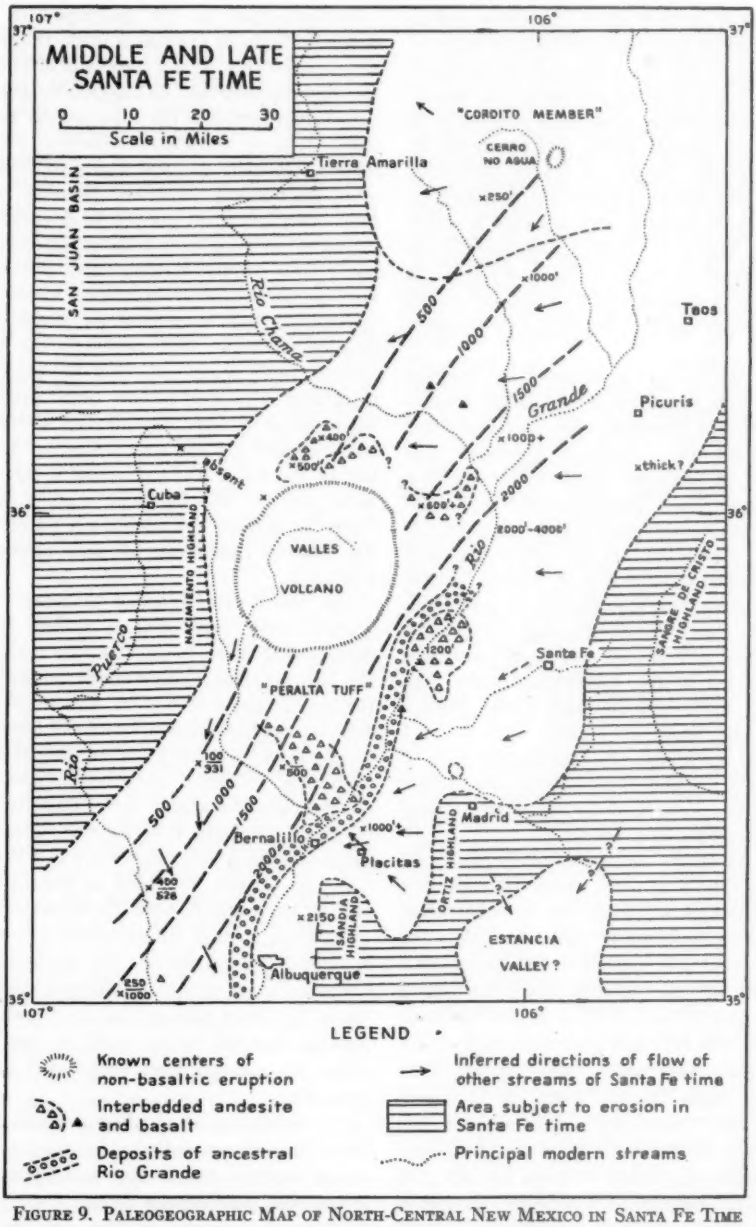

FIGURE 9. PALEOGEOGRAPHIC MAP OF NORTH-CENTRAL NEW MEXICO IN SANTA FE TIME

Domingo Valley (Bryan and Upson, unpublished manuscript) probably represents the same period of eruption and is latest Santa Fe,

(d) Contemporary eruptions of andesitic and basaltic lavas, principally in four areas peripheral to the Valles Mountains. Lavas in

two of these areas are interbedded with deposits of the ancestral Rio Grande.

(e) Accentuation of the Ortiz and Sandia highland areas, with incorporation of a part of the latter into the basin of deposition. At Placitas in particular, basal beds of the Santa Fe formation post-date Abiquiu (?) time by an interval during which northward drainage in the Tonque Valley was diverted to a westward drainage. By analogy with the Sierra Nacimiento, uplift of the Sandia block may be late Santa Fe; the subsequent drainage changes and eventual inception of deposition at Placitas are more surely late.

Near and south of the Valles Mountains, therefore, the geography sketched is late Santa Fe. The preceding and probably longer interval represented by the Middle Red member in the Albuquerque-Belen Basin may well have lacked the conspicuous volcanic accumulation in the Valles Mountains and the strong rejuvenation of highlands adjacent to the sedimentary basin. Northeast and north of the Valles Mountains, the sketch is based upon data from beds which can be dated only in the Española Valley. In this, the type area, abundant fossils imply that the bulk of the formation is older than the latest Santa Fe. These beds and those in the Abiquiu quadrangle farther north were deposited by streams flowing exclusively westward from the Sangre de Cristo Mountains to a master drainage west of any known exposures. There is no evidence of drainage northward from the Valles Mountains. Such a master drainage, of course, lay far west of the northernmost known deposits of the ancestral Rio Grande. Bryan (1938, p. 207, fig. 46) postulated an abrupt change in the course of the master stream, so that it flowed generally east-west across the northern portion of the present Valles Mountains. Such an abrupt swing might indeed arise from diversion of a through-flowing steam such as that postulated for Abiquiu time around an area of active uplift in the Sierra Nacimiento. One might suppose, however, that volcanic accumulations in the Valles Mountains, and peripheral to them, which nearly impinge upon the northern portion of the uplifted area, would have further diverted the master stream northward into the Abiquiu quadrangle. Denny (1940b, Pl. 1 and sections) has shown that the northernmost outcrops of river gravels are probably younger than the remainder of the Santa Fe formation in the Española Valley, which dips generally westward. Hence, river deposits may be absent in the Abiquiu quadrangle only because the beds now exposed are all older than the known river deposits farther south. It is tentatively suggested that the Santa Fe formation in the Española Valley and in the Abiquiu quadrangle is principally equivalent to the Middle Red member in the Albuquerque-Belen Basin, and that the master drainage of that interval lay some distance west of the position plotted in Figure 9. The known river deposits, on the other hand, are tentatively regarded as entirely of late Santa Fe age and later than the drainage changes induced by uplift of the Sierra Nacimiento and volcanic eruption in the Valles Mountains.

Butler (1946, p. 100–102) has shown that the Santa Fe formation in northern New Mexico grades laterally into volcanic detritus, named the Cordito member of the Los Pinos formation, deposited by streams flowing westward into the Tusas quadrangle (p. 85–86). The volcanic débris was probably contributed by contemporary centers of eruption in the Taos Plateau; one contemporary center is identified as Cerro No Agua (p. 77–78). Most of the other known volcanic centers of the Taos Plateau (not plotted in Fig. 8) lie too far south to account for the distribution of the Cordito member; additional centers must be buried beneath younger basalt flows northeast of Cerro No Agua. The interval of volcanic activity in northernmost New Mexico cannot be related to events in the Albuquerque Basin with assurance.

Modern Rio Grande Depression

The normal faults of the Galisteo-Tonque area, initiated in late Santa Fe time and probably continued after deposition ceased, are part of a complex system outlining the modern Rio Grande Depression. The general pattern of faulting in north-central New Mexico is illustrated by those areas in which the boundary faults have been mapped, although details are still lacking for much of the area. Within the Rio Grande Depression, the common lack of

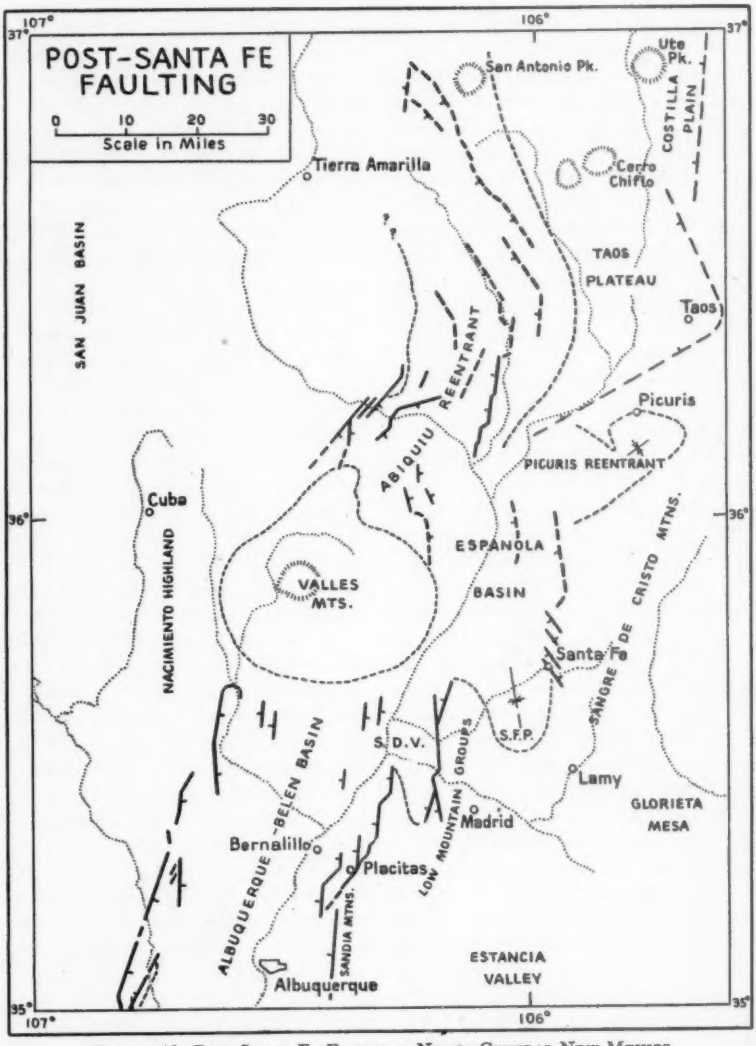

FIGURE 10. POST-SANTA FE FAULTS IN NORTH-CENTRAL NEW MEXICO
Compare with Figure 6.

horizon-markers within the several basin deposits and consequent incompleteness of structural data leads to a less complete picture of the regional deformation. The principal known faults, generalized in most instances, have been compiled from the sources referred to previously (Fig. 10).

Within the area shown, the Rio Grande Depression comprises several subdivisions, more or less defined by the fault pattern as so far known. Bryan (1938, p. 213), using the structural data then available, pointed out that the several subdivisions could be grouped into two north-south systems, the Española-Santa Fe Basin and the Albuquerque-Belen Basin, conspicuously offset between Bernalillo and

Santa Fe. The Abiquiu Re-entrant, north of the Valles Mountains, was considered as a local enlargement of the Española-Santa Fe Basin, interrupting the generally linear system. Butler (1946, Pl. 1, fig. 25) has since shown that the fault systems on the east side of the Abiquiu Re-entrant can be traced northward into the Tusas quadrangle, with a change of trend to the northwest. Although data are lacking further west, the pattern of these faults suggests that the Abiquiu Re-entrant is a semi-independent, linear depression, parallel and adjacent to the series of troughs tending northward from the Española Basin. The Abiquiu Re-entrant thus extended lies on a projection of, and continues the general trend of, the Albuquerque-Belen Basin farther south.

The Abiquiu Re-entrant is separated from the Albuquerque Basin by the great volcanic pile of the Valles Mountains. A part of this pile is late Santa Fe and post-Santa Fe, but an older volcanic series (Chicoma) is believed to be pre-Santa Fe (Smith, 1938, p. 939–940; 958). The older volcanic pile was apparently bevelled and incorporated in the late Tertiary sedimentary basins, there to be buried by the Abiquiu (?) and Santa Fe formations. Its present exposure implies that, relative to the areas north and south, the Valles Mountains are structurally high. The structural boundaries of this block are largely concealed by post-Santa Fe lavas, but faulting has been recognized locally (Smith, 1938, p. 961–962, figs. 4, 10; Bryan and Upson, unpublished manuscript).

The Rio Grande Depression in north-central New Mexico thus appears to comprise two parallel and adjacent series of basins. The westerly series is more regular in general trend and outline, but is interrupted in the Valles Mountains. The easterly series is highly irregular; individual basins are diverse in trend and in part of diverse ages. The Española-Santa Fe Basin and Picuris Re-entrant contain no basin deposits younger than the Santa Fe formation, but were deformed in post-Santa Fe time. The Taos Plateau, next to the north, is underlain by thick accumulations of basaltic lava and interbedded stream deposits which are post-Santa Fe in age, and the structural depression itself may be younger than those to the south. Just south of the Colorado line, the Taos Plateau gives way to a still younger depression, the Costilla Plains, in which fault scarps attest that movements have persisted to recent time. The easterly series of basins apparently dies out southward in a broad syncline in the Santa Fe Plateau. South of the area mapped, the modern Rio Grande Depression is continued by the broad, single trough of the Albuquerque-Belen Basin.

The faulting which outlined the modern Rio Grande Depression also led to transition from deposition to erosion in the structural basins of Santa Fe time. The subsequent history of the Galisteo-Tonque area has been dominated by erosion. Highlands remain in those areas underlain by resistant rocks; the remainder of the area was gradually reduced to a uniform erosion surface, the Ortiz pediment, later dissected by successively rejuvenated streams. Stream gravels associated with remnants of this surface conceal the older rocks in much of the area (Pl. 1). The general progress of erosion was interrupted by renewed volcanic activity in the late Santa Fe centers of the Cerros del Rio and the Mesa Santa Ana. It is proposed to describe the post-Santa Fe modelling of the present landscape in a subsequent paper.

Summary

Late Tertiary time was initiated by strong deformation in the Galisteo-Tonque area, accompanied by the intrusion of dikes, sills, laccoliths, and stocks. Principal intrusions are now exposed in the north-south line formed by the Cerrillos Hills, Ortiz, San Pedro, and South Mountains; dikes and sills are exposed in the Galisteo Lowland and the Tonque Valley as well. The structures produced lay athwart an early Tertiary basin of deposition and are interpreted as deformation in a block moving relatively northeast in a horizontal couple, marked by shear in the Tijeras and Los Angeles fault systems. Deformation may have been localized in an area adjacent to the fault systems. Contemporary structures have not been differentiated with certainty elsewhere in northern New Mexico. In fact, deformation of the widespread Potosi volcanic series has been ascribed chiefly to movements

associated with late Tertiary (post-Santa Fe) faulting (Butler, 1946, p. 162). Ray and Smith (1941, p. 206–207) have inferred that a widespread peneplain was developed in northern New Mexico during "mid-Tertiary" time.

It has been customary to assign the structural features of north-central New Mexico, younger than Cretaceous rocks and older than late Tertiary basin deposits, to an episode of compression called "Laramide". In the Galisteo-Tonque area, midway between the Sangre de Cristo Mountains and the Manzano Mountains, where steep thrusting occurs (Reiche, 1949, p. 1209–1210; Read *et al.*, 1944), one can distinguish three episodes of deformation. The first, early Eocene, is recorded by the varying thicknesses of upper Cretaceous rocks removed by pre-Galisteo erosion (Stearns, 1952a, fig. 1). The second, late Eocene, is shown by isopachs of the Galisteo formation (Stearns, 1943, fig. 6). The third, and most intense, Oligocene, is post-Espinaso and pre-Abiquiu (Fig. 7). The question thus arises: "What is Laramide?". There is no simple answer. The question must await elaboration and synthesis of data both from the southern Rocky Mountains and from central New Mexico. However, the question underscores the warning of Burbank and Goddard (1937, p. 976) that "Laramide" deformation is a complex history, not simply of renewed deformation on old trend lines, but of deformation patterns changing in trend as well as in intensity.

An Oligocene interval of erosion, during which the Cieneguilla limburgite was extruded locally, preceded initiation of the Abiquiu basin of deposition. This basin was developed by downwarping of two broad, adjacent areas. The northern area was filled principally by volcanic detritus from contemporary centers of eruption. Volcanic detritus spread southward into the southern basin, but it was principally filled with nonvolcanic débris from adjacent highland areas. It is inferred that a major stream entered the southern basin from the west, flowing across the present site of the southern Valles Mountains. It may have continued through and out of the basin, west of the Sandia Mountains, but on this point further data are needed from the east side of the Albuquerque-Belen Basin. The southeast margin of the basin was irregular, because highlands raised by earlier deformation had not been completely reduced by erosion.

The younger Santa Fe basin of deposition appears to have been an asymmetric trough, with maximum deposition near the eastern margin. The Albuquerque-Belen Basin, as far north as the Galisteo-Tonque area, was probably bounded to the east by strong faults. Farther north, the eastern limits of deposition may have been fixed only by warping. The area of deposition corresponded generally to that of Abiquiu time, but there are many differences in detail. In "middle Santa Fe" time, the drainage pattern may have been similar to that of Abiquiu time, except that the Abiquiu tuff fan was buried by granitic detritus from the Sangre de Cristo Mountains, east of the volcanic centers. By late Santa Fe time, the earlier major drainage entering the basin from the west must have been diverted or disrupted by uplift of the Nacimiento Highland and construction of the Valles Volcano. The ancestral Rio Grande may represent that drainage, diverted north of the new highland areas into the Abiquiu Re-entrant, then flowing southward in the trough between fans built eastward from the Valles Volcano and westward from the Sangre de Cristo Mountains. Concurrently, the former northward drainage from the Tonque Valley was diverted westward across the north end of the early Sandia highland. Renewed uplift of the Sandia Mountains gave rise to northward drainage in the Placitas area which alternated with, but never replaced, westward drainage from the Ortiz highland.

Marginal faulting, initiated in late Santa Fe time, outlined the modern Rio Grande Depression and, in much of north-central New Mexico, led to a change from deposition to erosion in the structural basins. The younger structural depression is for the most part bounded by faults lying well within the margins of the earlier depositional basins. In north-central New Mexico, the depression is double. The western portion, including the Albuquerque-Belen Basin and the Abiquiu Re-entrant, is a true graben, interrupted by the structurally high Valles Mountains. The eastern portion is an irregular series of basins of diverse age and trend, outlined in part by warping and in part by faulting. The northernmost basins have been actively in-filled during post-Santa Fe time,

and appear to be asymmetric troughs bounded by faults along the east margin. The Picuris Re-entrant, and the Española Basin contain only beds of Santa Fe age and appear to be the product of flexure as much as of faulting. The eastern basin system appears to die out in the Galisteo-Tonque area; farther south, the Rio Grande Depression is continued by a single graben, the Albuquerque-Belen Basin.

Appendix. Log of North Albuquerque Acres Well No. 2 ("Norins Well")

(Drilled by Norins Realty Co.; T 11 N. R. 4 E., Sec. 19; quoted from well log 1486, New Mexico State Bureau of Mines and Mineral Resources.)

Drilled with cable tools

Formation	Bottom, feet
Surface soil	6
Gravel	20
Sand	26
Boulders	34
Coarse sand	78
Boulders	100
Fine sand	138
Boulders	150
Cemented gravel; hard	160
Sand and gravel; softer	205
Cemented granite; arkose	210
Soft sand	220
Arkose	225
Soft sand	230
Arkose	250

Changing to rotary

Arkose; very hard	350
Soft sand	360
Granite wash; very hard	390
Granite wash; hard	415
Soft sand	442
Granite wash; hard	450
Coarse gravel	475
Arkose	500
Arkose; very hard	550
Arkose; very hard with thin strata of red shale	600
Gray and red shale	620
Hard coarse cemented gravel	670
Red and gray shale	700
Arkose; hard	750
Gray shale	760
White hard sand	800
Arkose; very hard	820
Arkose; very hard, with red and gray shale breaks	850
Red and gray shale	950
Coarse sand and gravel carrying water	960
Brown shale	1000
Quicksand	1050
Sand and gravel	1100
Brown shale	1150
Soft sand	1175
Coarse gravel	1200
White lime	1210
Brown shale	1250
Sand and gravel	1300
Brown shale	1310
Coarse gravel	1325
Sand	1375
Coarse gravel	1425
Brown shale	1460
Gray lime	1473
Fine sand	1550
Gray and red shale	1600
Sand	1625
Coarse gravel	1675
Granite boulders	1690
Hard coarse granite wash	1700
Hard granite wash	1750
Soft sand	1775
Boulders	1800
Granite wash	1850
Red shale	1875
Gray shale	1925
Granite wash	1935
Granite boulders	1950
Brown shale	1960
Hard lime	1975
Gray lime	2000
Red shale	2025
Hard granite boulders	2050
Granite boulders; hard	2100
Granite wash	2150
Gray shale	2160
Hard gray sand	2180
Gray shale	2345
Gray sand	2400
Gray sand, hard	2410
Gray shale, soft	2500
Soft gray shale	2550
Hard gray shale	2600
Hard gray sand	2650
Hard gray sandstone	2700
Hard gray sand	2750
Gray sand, softer	2800
Gray sand, hard	2850
Gray shale	2890

Formation	Bottom, feet
Gray soft sand; water	3002
Gray soft sand	3110
Soft sand	3118
Gray shale	3218
Very hard sand, gray	3300
Gray sand; softer	3350
Soft sand	3400
Very hard gray sandy lime	3445
Hard gray sandy lime	3480
Gray shale	3570
Lime shells	3600
Pink shale	3640
Gray sand, hard	3650
Gray sand	3700
Gray sand, hard	3780
Gray shale	3810
Sand, gray, soft	3820
Gray sand, harder	3900
Gray sand, hard	3960
Gray sand, softer	4020
Brown shale	4040
Gray sand, hard	4100
Black sand, soft	4300
Black and gray sand, harder	4400
Gray sand, hard	4460
Dark gray sand, hard	4480
Gray sand, hard	4560
Hard gray sand	4600
Gray sticky shale	4637
Gray sticky shale	4710
Gray hard sand	4776
Gray sticky shale	4780
Gray hard sand	4795
Gray shale	4810
Gray hard sand	4828
Gray hard sand with strata of sticky shale	4870
Gray shale	4930
Gray hard sand	4950
Gray soft sand	4970
Gray sticky shale	4980
Gray sand, soft	5024
	T.D.

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